The Late Devensian Deglaciation in the Cairngorm Mountains, Scotland

Jeremy Everest MA (Edin)

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Declaration

I made this thesis. All the work included within these pages- text, maps and figures- are original and my own, unless indicated otherwise.

"...when he found the Truth he sought no further; but from that day forth,
with his soldering-iron in one hand, and his bludgeon in the other,
he tinkered its leaks, and reasoned with objectors." Mark Twain

"Better toil blindly, beating every stone in turn for grains of gold,
whether they contain any or not, than lie down in apathetic decay." John Muir

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I would like to thank Professor David Sugden for his unfailing help over the last six years. I have benefited from his unsurpassed knowledge of the Cairngorms, his insight and critical view of my wordy writing style. Great thanks must also go to Dr Andy Dugmore for his boundless enthusiasm and ability to link the most disparate evidence to form convincing theories. Dr Bill Phillips has also been on the receiving end of many mathematical and isotopic physics enquiries, and without whose help my O level standard understanding of these areas of my research simply would never have been enough. Dr Peter Kubik at ETH in Zurich has also given invaluable aid in the derivation of cosmogenic ages from AMS results, as well as running all my samples. Alan Davidson has had the unenviable task of doing much of the 'dirty work' in the lab, and without his help the thesis would have neared completion sometime in the next decade. My research has been funded by the Department of Geography, the Carnegie Trust for the Universities of Scotland and the Natural Environment Research Council.

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Finally Morag Dickson has been my light. I cannot thank her enough, but she knows I hope.

Abstract

The aim of this thesis is to use cosmogenic isotope analysis to date a lateglacial stillstand in the eastern Highlands of Scotland. The dates link the extent of the local Cairngorm ice cap and the Scottish Ice Sheet at a particular stage of deglaciation. The constraints provided by such dates add insight to the process of deglaciation and the climatic conditions at the time. The geomorphological and sedimentological evidence suggests that the stillstand lasted around 1 ka and was the last of two or three such stages that affected the region. Cosmogenic ^{10}Be results show that the stillstand occurred at 16 - 15 ka BP and was synchronous across the massif. Correlations between the Cairngorm ^{10}Be ages, data provided by North Atlantic ocean cores, and the Greenland ice core data suggest that the stillstand in the Cairngorms occurred at the same time as Heinrich 1, the final major iceberg rafting event in the North Atlantic at the end of the last glaciation. Independent ^{14}C and tephrochronological dating from Loch Etteridge within the Spey Valley show that deglaciation proceeded rapidly after the stillstand. This latter data has been used to provide constraints on erosion rates of granite surfaces in the Cairngorms which are less than 10 mm / ka.

Two further implications arise from this work. First the last stage of valley glaciation in the Cairngorms occurred at 16 - 15 ka BP, and thus any Loch Lomond Stadial event (Younger Dryas equivalent) in the Cairngorms was confined to the high corries. Second, the dating of the stillstand shows that the eastern flank of the main Scottish Ice Sheet had retreated to 40% of its LGM extent by the time of the Heinrich 1 Event.

Chapter 1:

Introduction and Background

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1.1 Aim & Introduction

This study is concerned with the use of landforms created by glaciers to learn more about the linkages between ice sheet systems and climatic change. The specific aim of the thesis is to establish a deglacial chronology for the western Cairngorms during the Lateglacial Interglacial Transition (LGIT), using cosmogenic ^{10}Be, and place this is in a North Atlantic context.

The Quaternary environment of the North Atlantic is an area and period that has received intensive study. The amount of high resolution data that has been derived from the region in the last thirty years is impressive. Through careful synthesis of this data, workers have begun to piece together an environmental history of the region. Although there are, and will always be arguments over the precise interactions of ocean, atmosphere and ice sheet systems, a broad-scale chronological framework for the Lateglacial period is emerging. This chapter draws on the records from the region, and attempts to fit ice sheet behaviour in the Cairngorm mountains into the broader North Atlantic context. The most straightforward way of achieving this aim is to deal with each of the main forcing mechanisms for glacial change in Scotland in turn; i.e. those of ocean circulation, atmospheric motion and continental ice sheet behaviour. From summaries of each, the chapter will assess Scottish deglaciation from a countrywide to finally a Cairngorm perspective.

It is generally assumed that glaciers are useful indicators of climate change, in that they respond to variations in temperature and precipitation on scales of decades or less. Changes in glacier flow, extent and thickness may be identified in the geomorphological record by the study of suites of landforms and sedimentary sequences, moraines, trimlines, and erratic boulders. Relative chronologies of deglaciation may be constructed from geomorphological relationships which demonstrate a sequence of change over time. Before the invention of various dating techniques, many studies were conducted successfully employing interpretation of landforms across large areas of deglaciated landscapes (Hinxman and Anderson, 1915, Charlesworth, 1956, and others). Modern studies now employ a range of dating techniques, such as ^{14}C, thermoluminescence, tephrochronology and cosmogenic surface exposure dating in order to place reconstructions of deglaciated landscapes into firmer chronological contexts. By correlating data collected from these studies with palaeoenvironmental data found in ice and marine cores, glacial geomorphologists can now link the pattern of deglaciation into a longer history of climatic and environmental change.

1.2 The North Atlantic Region following the LGM

1.2.1:Introduction

During the last glacial period (70 - 10 ka BP) the circulation of the North Atlantic had a profound effect upon the growth and decay of the ice sheet systems surrounding its margins, namely the Laurentide Ice Sheet (LIS) based on continental North America; the Greenland Ice Sheet (GIS); the Fennoscandinavian Ice sheet (FIS) and the British Ice Sheet. Through changes in Sea Surface Temperature, Density and Salinity (SST, SSD and SSS) and the movement and location of oceanic currents, the transportation of heat from the Equator to the Poles is modified (Figure 1.1). Recent studies have attempted to create reconstructions of climatic and oceanic change at resolutions of less than 1000 years (Lowe *et al.* 1995, Hoek & Bohncke, 2001, Hald *et al.* 2001 and others). Most ice core records show that the last Glacial period was climatically extremely unstable (Dansgaard *et al.* 1982), characterised by rapid climatic transitions, 'Dansgaard- Oeschger' events, abrupt changes leading to cold periods of hundreds to thousands of years *(Bond et al. 1993).*

1.2.2: The Conveyor

Until recently the governing paradigm of oceanic flow during the last glacial has been that of the 'Thermohaline Superconveyor' system (Broecker and Denton, 1990a, 1990b, Broecker *et al.* 1989, Fanning & Weaver, 1997, Bond *et al.* 1993, Lehman & Keigwin, 1992 and others). Northward flow of warm, saline surface water transfers a large amount of heat to the atmosphere on cooling. As it cools it mixes with surface water from the Arctic and sinks to create a deep convection current (North Atlantic Deep Water or NADW, Figures 1.1 and 1.2). When this system breaks down, poleward transport of warm water is halted, and glacial conditions predominate in the North Atlantic region. The model proposed by many authors, though predominantly by Broecker and Denton (1990b), suggests that circulation may oscillate between 'off' and 'on' modes through the effect of changing ocean heat flux on the melting rates of ice, both terrestrial and sea ice, situated near to the convecting system. Periods of strong thermohaline flow lead to accelerated melting of the former mid-latitude ice sheets. These huge meltwater events are termed Heinrich events, and are characterised by massive iceberg calving.

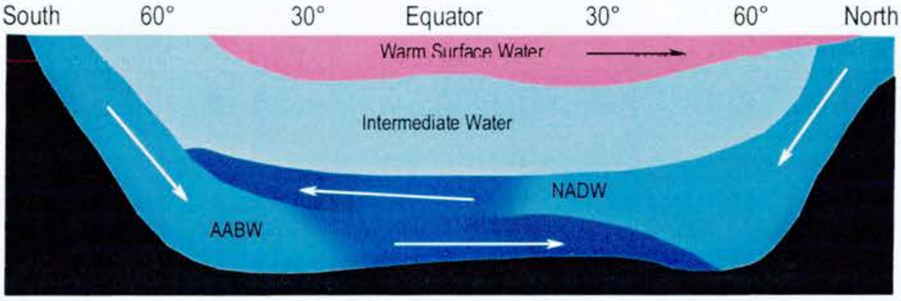


Figure 1.1: NADW formation. Warm water flow in the North Atlantic cools and sinks in the Polar region, forming low - salinity NADW, the 'pump' driving the Conveyor system.

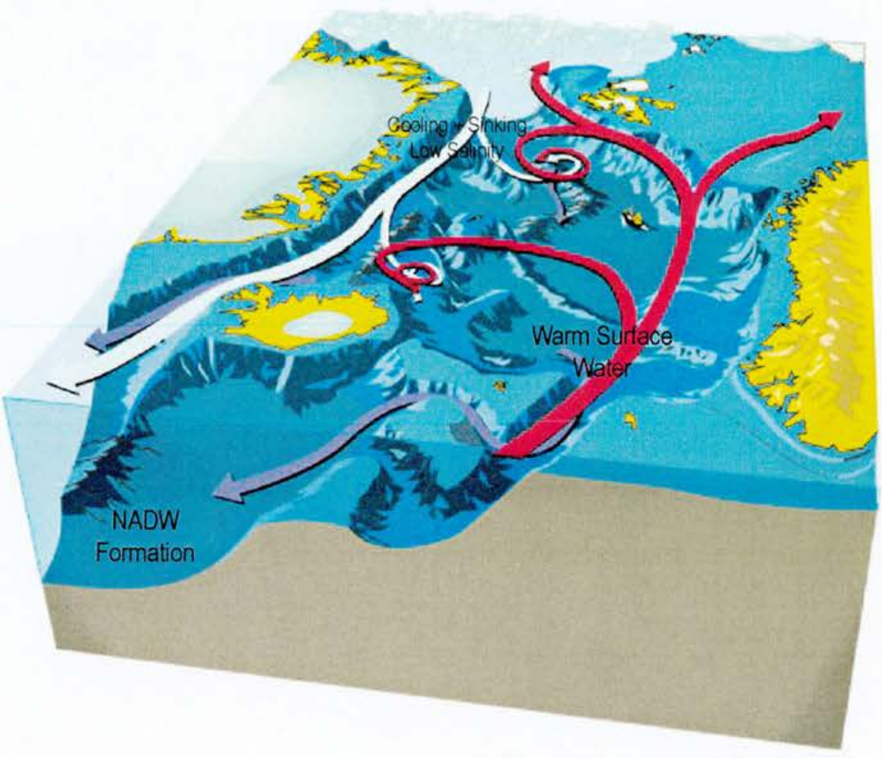


Figure 1.2: NADW formation in the Fram Strait and Norwegian Sea. Flow is channelled at depth by bed topography, including the Rockall Trough, and Iceland and Irminger Basins (from NOAA website).

These events are visible in the marine sedimentary record, with distinct bands of Ice Rafted Debris (IRD) and changes in the isotopic content of sediment indicating a large influx of cold, nutrient poor and low salinity meltwater. The increased meltwater then may damp and halt the overturning oceanic circulation, and the production of NADW. The cooling that results then halts melting of the ice sheets, and salinity of the surface waters then increases to levels high enough to restart the conveyor system. The behaviour of this system has led authors to describe it as the 'Flickering Switch' of climate change (Taylor *et al.* 1993) and it appears to be a reasonable explanation for at least some of the rapid climatic transitions that characterise Dansgaard - Oeschger events. A further development of the conveyor' hypothesis is that the Younger Dryas period, generally accepted to have occurred between 12.4-12.6 and 11.3-11.5 ka BP, was caused by the diversion of meltwater drainage from the Mississippi drainage system to that of the St Lawrence river. This would have caused an influx of cold water to the northern part of the western Atlantic, leading to cessation of northward flow, and NADW production. The 'conveyor' hypothesis provides a concise and testable theory upon which we can base our understanding of oceanic processes over the last 70 ka.

A development of the 'conveyor' theory has been proposed by many authors more recently (Sarnthein *et al.* 1994, 1995, Veum *et al.* 1992, de Vernal *et al.* 1996, Maslin *et al.* 1995, Lehman & Keigwin, 1992, Lehman *et al.* 1991). Most agree that the 'conveyor' exists and is the primary cause for fluctuations in climate in the North Atlantic region. However most propose that the response to these fluctuations is far more complex than earlier theories suggest. Boyle (1988) and Veum *et al.* (1992) identify nutrient-depleted intermediate waters in the North Atlantic during the LGM which could be responsible for the drawdown of CO_2. The source and origin of these waters has been discussed at length, and the most likely area for their formation is the area occupied by the Greenland, Iceland and Norwegian Seas (GIN). The GIN seas contribute to deepwater today, but Veum *et al.* suggest that during LGM conditions they formed intermediate waters instead. Such a situation would lead to convection and brine formation in the higher latitudes of the North Atlantic. This overturning to intermediate depths at higher latitudes would deplete the upper ocean of nutrients, and increase nutrients in the deep water system (Veum *et al.* 1992). Boyle (Boyle, 1988, Boyle and Rosener, 1990) suggested that this mechanism would be the main cause of the 30% reduction in CO_2 content of the atmosphere during the LGM. This nutrient depletion in the North Atlantic may have led to the increased uptake of CO_2 in other regions of the world ocean. The main difference that authors such as Veum and Sarnthein introduce with regard to the superconveyor theory is the operation of the conveyor during the deglacial period. Before Sarnthein (1994, 1995) had produced his theory Veum (1992) argued that during the final stages of deglaciation strong warming occurred in the Norwegian Sea at 15.5 ka BP (Lehman & Keigwin, 1992). This implies that melting and deglaciation occurred during times of reduced thermohaline circulation, i.e. when the conveyor was in 'off' mode. Evidence shows that NADW formation restarted at 14.1 ka BP. A major pulse of the northern hemisphere ice masses (the LIS and GIS in particular) occurred at approximately 14.2 ka BP, and therefore may have been caused by heat released from the conveyor. The European warming happened before this, at a time when the conveyor is thought to have been weak, or in 'off' mode. The implication is that warming therefore must have taken place at times of both 'on' and 'off' modes of conveyor circulation.

The Younger Dryas period provides further evidence for the non-linear mode of operation of the ocean atmosphere system during deglaciation. Evidence shows that a modern conveyor system operated throughout the Younger Dryas (Veum *et al.* 1992). Although the Younger Dryas was a return to glacial climates in northern Europe, and to glacial sea surface temperatures, it was not a return to a glacial ocean circulation. Neither does the theory of the diversion of meltwater drainage from the LIS, from the Mississippi to the St, Lawrence as the cause of the Younger Dryas stand up in the light of new evidence. The changes in salinity caused by this diversion were small, and certainly not of the order of magnitude needed to halt the conveyor (de Vernal *et al.* 1996). Indeed the salinity values for the continental shelf off the St. Lawrence are very similar, if not lower than today's, indicating reduced meltwater runoff before and during the Younger Dryas. Therefore it can be argued that a meltwater pulse of sufficient size into the western north Atlantic never occurred, and the causes of the Younger Dryas must be sought elsewhere.

The most significant new view has been proposed by Sarnthein *et al.* (1994, 1995). They suggest that there are three major modes of oceanic circulation operating during the deglacial phase. The first is the Holocene or Interglacial Mode, where strong southward-bound NADW flows in the Northeast Atlantic, originating in the Norwegian Greenland Sea (Figure 1.3). This mode also applies to the Younger Dryas. The next mode is the Glacial Mode. NADW still flows southward but at a much

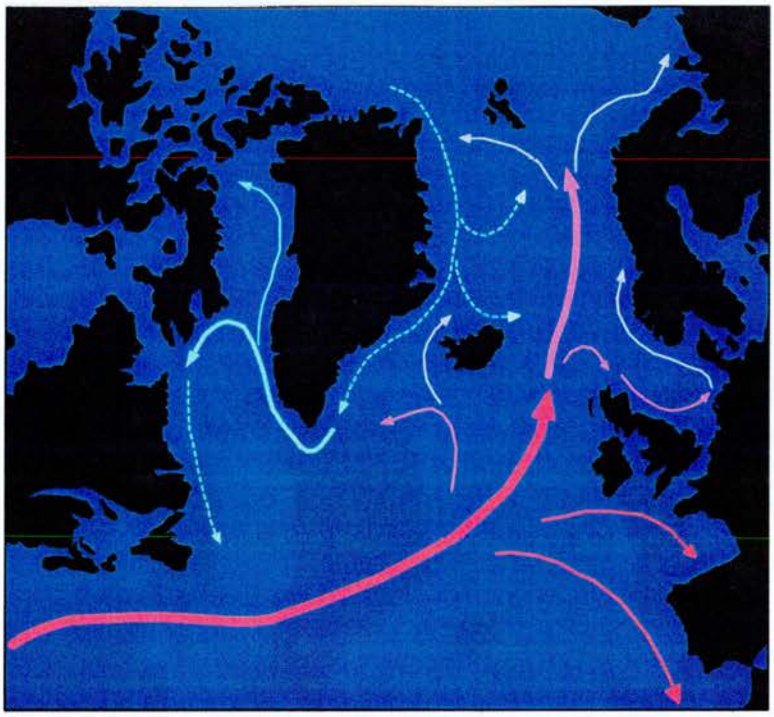


Figure 1.3: Modern or 'Interglacial' surface water flow in the North Atlantic region (Red- warm current; purple- cooling; blue - cold. Thickness of line denotes strength of current. Dotted lines indicate weak flow) (after Sarnthein, 1994, 1995).

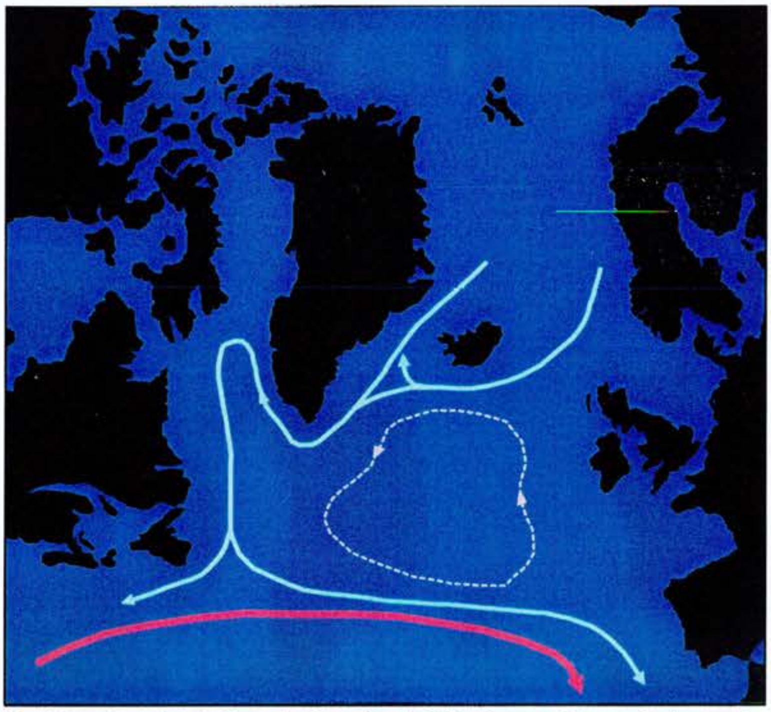


Figure 1.4: North Atlantic circulation during the LGM. Oceanic and atmospheric polar fronts have moved south of the European mainland, abutting the coast of Portugal, and diverting warm water flow away from northern Europe. North Atlantic surface water flow is therefore dominated by cold water flowing from polar regions (after Sarnthein, 1994).

reduced volume, although its area of formation shifts to south of Iceland (Figure 1.4). Finally the Meltwater Mode occurred during major meltwater events, the first between 13.8 ka BP and 12.8 ka BP when a major meltwater pulse originated from the catastrophic break up of the Barents Sea Ice shelf, and the second immediately after the termination of the Younger Dryas, between 12.2 ka and 11 ka BP. Such was the severity of this halt to poleward heat transport that temperatures in the region could have been anything up to 14° C lower than during the Holocene (Siegert, 2001).

1.2.3: Dansgaard-Oeschger and Heinrich Events

Within the time period of the last glaciation, various lines of evidence show a cyclical system of climate change. Greenland ice cores and marine sediment core records from the North Atlantic display a number of periods occurring every 2 or 3 ka during the cold period. These fluctuations are of sub-Milankovitch periodicity, and are characterised by a period of gradual cooling, followed by a rapid warming phase. Associated with these temperature variations an increase in ice accumulation rates in Greenland ice cores, an increase in dust concentrations and in the concentrations of methane and CO_2 (Broecker, 1994). These cycles have been termed Dansgaard-Oeschger events.

At least six of these events have culminated in the probable surging of the Laurentide Ice Sheet in North America, producing vast numbers of icebergs in the North Atlantic. Such iceberg 'armadas' are termed Heinrich events (Heinrich, 1988, Broecker, 1994), and have been linked to ice sheet advance in several locations (Bowen *et al.* 2002, Clark *et al.* 1999, Grousset, 2000). Not all D-O events have led to Heinrich events. Bond *et al.* (1993) have shown that the Heinrich events occurred during the coldest phase of the last glacial period, and that each Heinrich event led to the initiation of a new series of D-O cycles. Current thinking argues for the Hudson Strait as the prime source of icebergs during Heinrich Events, and thus the major source of IRD (Andrews, 1998). Other ice sheet systems may have delivered quantities of IRD to the North Atlantic, but their role is less significant than that of the Hudson Strait ice stream (Dowdeswell, *et al.* 1999). One recent theory is that surging of the European ice sheets led to eustatic sea level rise, initiating rapid calving of the marine margins of the Laurentide ice sheet system. Evidence for this comes from marine sedimentological data, indicating European provenance of IRD at the beginning of some Heinrich events (Grousset *et al.* 2000). Icebergs calved from a terrestrial ice sheet margin are inevitably laden with glacial debris, and as they float across the ocean, this debris melts out and accumulates on the sea bed. Six layers of IRD have been positively identified in the North Atlantic, with at least two of these possibly containing at least two pulses of debris accumulation, pointing to a pulsed sequence of rafting events (Lebreiro *et al.* 1996). The massive input of a source of fresh, cold water to the surface of the North Atlantic had the effect of reducing surface salinity, and possibly halting normal oceanic flow, disrupting the conveyor, and reducing the transfer of heat polewards.

The timing of H events has been linked to abrupt climatic changes in the North Atlantic region (Broecker, 1994). H1 marks the onset of the termination of the last glacial period, occurring prior to the Younger Dryas; H6 is thought to have occurred at the transition from the last interglacial to the cold conditions in the North Atlantic at the commencement of the last glaciation (Figure 1.5). Unfortunately the resolution of dating control from marine sediments provides problems for the correlation of accurate dates from IRD layers throughout the North Atlantic basin. The variety of source areas for marine cores, and variations in accumulation rates and bioturbation have led to disagreement over the precise timing of Heinrich events. Table 1.1 gives an indication of the range of ages for H events. Andrews (1998) argues that dates of H events, gathered from marine cores, are potentially unreliable for several reasons. Low sediment accumulation rates in areas distal from calving margins may allow bioturbation to affect data; too few ^{14}C ages have been used to bracket H events at most sites; finally the temporal and spatial variability of the marine ^{14}C reservoir may also affect correlation of stratigraphies across ocean basins.

Reference	Radiocarbon Age of H1 (ka BP)	Calendar Age of H1 (ka BP)	Mean Calendar Age (ka)	Age Range (ka)
Bond *et al.* (1992)	13.5 - 14.6	16.2 - 17.5	16.85	1.3
Elliot *et al.* (2002)	13.2 - 15.1	15.9 - 18.1	17	2.2
Broecker (1994)	14.3 - 14.5	17.1 - 17.4	17.25	0.3
Bowen *et al.* (2001)		16 - 17.2	16.6	1.2
Bond *et al.* (1993)	13.6	16.3	16.3	0
Andrews (1998)	13.2 - 14.6	15.9 - 17.5	16.7	1.6
Keigwin & Lehman (1994)	14.3 - 15	17.1 -17.9	17.5	0.8

Table 1.1: Published ages for Heinrich Event 1 in ka BP.

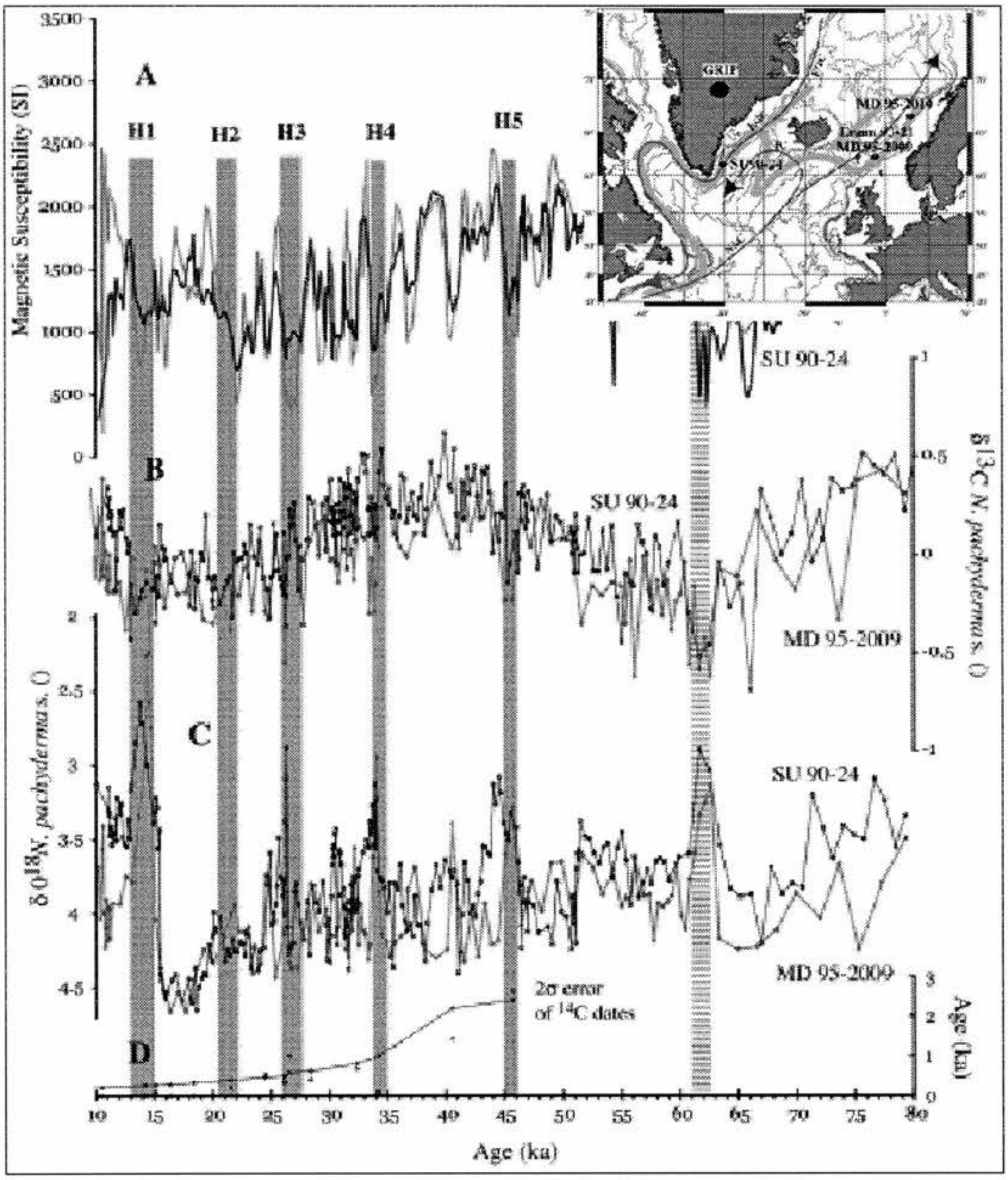


Fig 1.5: Ages of Heinrich events in North Atlantic cores (Locations of SU 90-24, MD 95-2009 shown in inset) in comparison to magnetic susceptibility of sediments, and δ O^{18} in N. pachyderma s. Note the spread on H1 age of between 15 and 13 ^{14}C ka BP (From Elliot et al. 2001).

Inevitably terrestrial ice sheet systems surrounding a cooling ocean will have responded. Recent data from Ireland shows a phased relationship between the British Ice Sheet and the collapse of the Laurentide Ice Sheet, with a readvance of the terrestrial ice cover in southern Ireland advancing at around 14 ka ^{14}C BP (Bowen *et al.* 2002). They argue that this readvance may correlate with Heinrich Event 1. Similarly Elliot *et al.* (2001) argue that the Fennoscandian Ice Sheet was able to respond rapidly to fluctuating atmospheric temperatures and moisture supply, caused by H events affecting the Norwegian Sea.

1.2.4: Heinrich Event 1

Evidence has shown that the meltwater pulse identified in the Atlantic by Broecker (1989, 1990a, 1994 and others) originated not with the collapse of the Laurentide Ice Sheet, but from the start of melting in the Norwegian, Greenland and Icelandic Seas, and by massive surging and collapse of the Barents Ice Shelf. Maslin *et al.* (1995) place a date of 17.1 ka BP on the peak of this melting event, and state that this Heinrich Event 1 had a greater effect on oceanic cooling and ice sheet collapse than all other Heinrich events. The exact date of the H1 event is unclear, as different core sites produce varying ages (Table 1.1); however the general agreement is that the H1 event took place between 17.1 and 14.4 ka BP. Rasmussen *et al.* state (1996a, 1996b, 1996c, 1997) that between 17.1 and *14.4-15.1* ka BP there was no convection in the Norwegian sea, in direct contrast to Sarnthein, who states that the conveyor system is still active at this time, though at a greatly reduced volume.

1.2.5: North Atlantic Palaeoclimate: Ice Core Records

Information derived from marine sources is extremely valuable in attempting to reconstruct palaeoclimates, however further high resolution information is available from ice core records. In themselves annual snowfall records may provide information concerned with variation in levels of precipitation over an ice sheet, though gases trapped within the ice provide further insights into the operation of climate over very long timescales. The GISP and GRIP cores from Greenland reveal climate records of over 100,000 years, and differences in the isotopic components within annual layers has provided proxy temperature, precipitation and atmospheric circulation data at a resolution that was inconceivable only a few decades ago.

1.2.5a: Palaeoclimate of the North Atlantic from Oxygen Isotopes

The use of $\delta\ ^{18}O$ as a proxy for temperature is well documented (Dansgaard, 1961, Dansgaard *et al.* 1982, 1993, Dansgaard & Tauber, 1969 and others). As temperature is depressed, the propensity for heavier isotopes of Oxygen to be transferred from the gaseous to the liquid state within water molecules is increased. Therefore ice layers formed from snowfall precipitated under colder conditions contains a relatively higher proportion of ^{18}O than ^{16}O. This variation can be seen in the Greenland ice core records, at extremely high resolution, stretching back well beyond the last glacial transition (Figure 1.6). Due to the relatively rapid fluctuations in atmospheric motion compared with oceanic flows (for example predicted temperatures in Greenland vary by ± 5°C within 30 years during deglaciation (Dahl-Jenssen *et al.* 1998)), it is not always possible to reconstruct broad scale atmospheric patterns with confidence. Tightly constrained reconstructions such as CLIMAP have been employed to better understand the role of these atmospheric reorganisations, and with increased complexity, linkages can now be made to oceanic systems.

1.2.5b: CLIMAP reconstruction of LGM Atmospheric Organisation

The CLIMAP reconstruction outlined the following situation in the northern hemisphere during the LGM:

1. Evaporation and precipitation were on average 15% lower than present July values.
2. Atmospheric circulation was dominated by enhanced east-west air flow, and a reduction in north-south circulation, thus reducing the amount of warm air transported northwards over the Atlantic.
3. The major land masses of Eurasia, North America and Greenland, covered by ice sheet systems, endured temperatures 20-30°C lower than present day values.
4. Very dry anticyclonic systems existed over the arctic ocean, as a result of permanent sea ice cover.

(From Siegert, 2001)

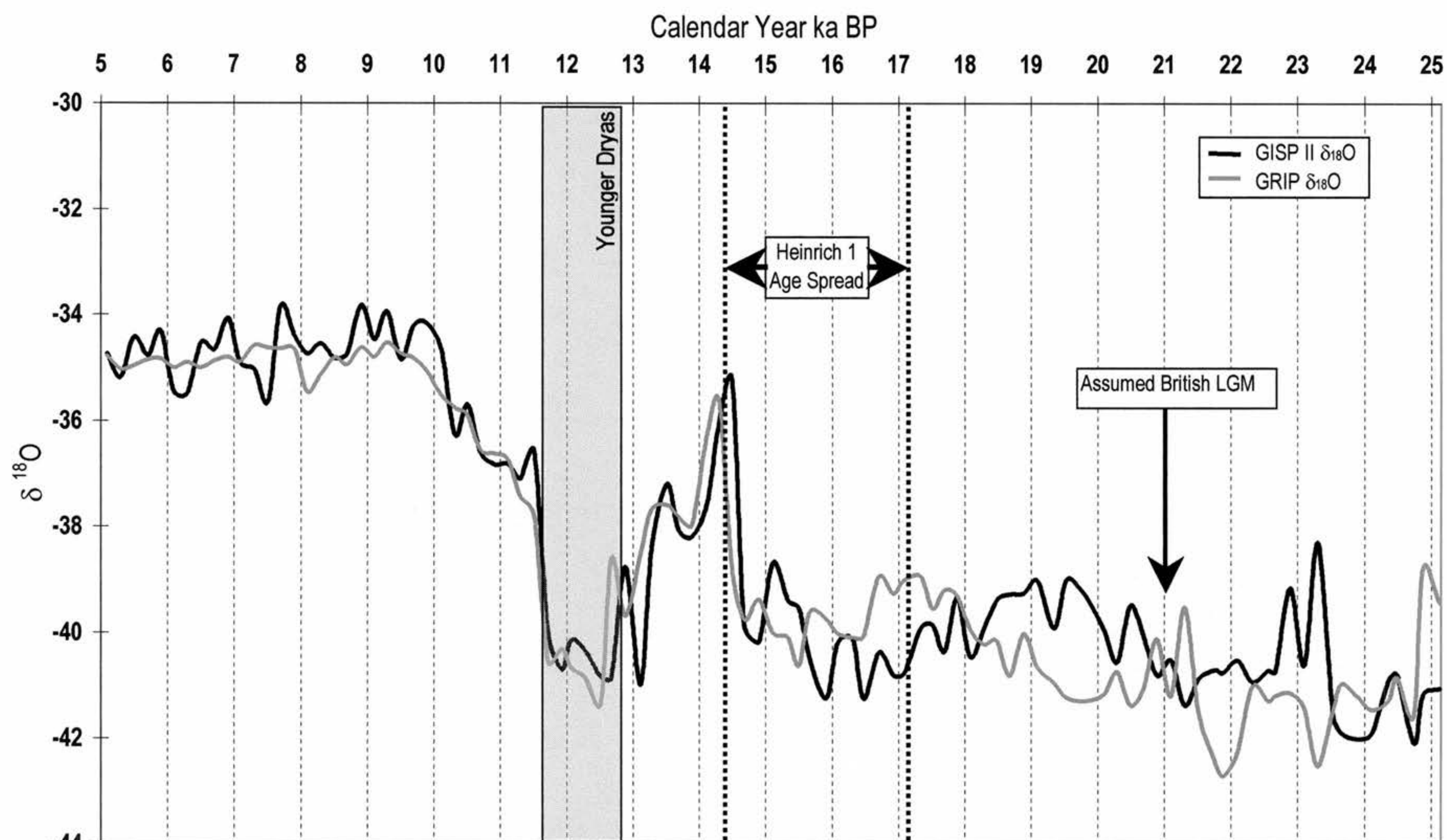


Figure 1.6: GISP2 and GRIP ice core proxy temperature records (note ages are Calendar years, not radiocarbon years).Timing of the Younger Dryas and Heinrich 1 iceberg rafting event, are shown.

Throughout the deglacial phase it is argued that the oceanic and polar fronts migrated northwards (Ruddiman and Macintyre, 1981), with a readvance southwards during the Younger Dryas. Associated with this northward movement is increased meridional transport of warm air from south to north in the North Atlantic, eventually leading to the termination of the last glacial at around 15 ka BP.

1.3: Chronology

1.3.1:The LGM

Most authors are in agreement that the LGM occurred at around 21 ka -17.3 ka BP. This period was characterised by a 130m fall in eustatic sea level (Jones & Keigwin, 1988), the build up of ice caps surrounding the Barents and North seas, and sea ice cover over the Norwegian, Icelandic, Barents and North Seas where they were still in existence, the North Sea being practically dry at this time (Figure 1.7).

The north east Atlantic was dominated by cold, low salinity, poorly ventilated waters. Sarnthein (1995) suggests that there was only seasonal ice cover in the Norwegian and Greenland Seas, and that the circulation regime was dominated by an Atlantic 'Palaeogyre' moving clockwise, and flowing southward west of Ireland. At such a time there was no east - west flow due to limited sea depth. Sea ice margins were in the Norwegian, Iceland and Greenland seas, and westward moving icebergs from these margins effectively halted warm water flow further east. Lehman *et al.* (1991) appear to be in part agreement with Sarnthein's observation of ice free Nordic seas, as they postulate that deglaciation of the Norwegian Sea began at around the LGM.

1.3.2: 18.5 - 17.1 ka BP

Several authors identify the start of a deglacial phase at around 18.5 - 15.5 ka BP inferred from δO^{18} signatures in marine cores from the North Atlantic (Duplessy *et al.* 1986, Lehman *et al.* 1991, Jansen & Veum, 1990, Jones & Keigwin, 1988, Bard *et al.* 1987). At 18.7 ka BP Lehman *et al.* (1991) propose synchronous ice shelf retreat of both the Barents and Fennoscandinavian Ice Shelves (Figure 1.8). This is supported by the work of Jones and Keigwin (1988) who state that a 15 m eustatic sea level rise, observed in the Bahamas (Fairbanks, 1989) is attributable to the catastrophic melting of the Barents Sea Ice Shelf. This melting continued until around 16.8 ka BP, and produced large amounts of cold surface water and icebergs that travelled westward through Fram Strait and into the North East Atlantic.

Figure 1.7: (after Anderson and Borns, 1994. These reconstructions have been chosen as they illustrate a generalised picture of the North Atlantic region throughout deglacial phase. Other reconstructions differ at a local scale eg. Ehlers & Wingfield, 1991) The North Atlantic region during LGM conditions. Sea ice spread as far south as the northern coast of Spain, and major ice masses covered almost all of the northern part of the European continent. Sea levels were around 120 m lower than today, consequently the British Isles formed part of mainland Europe

Figure 1.8: The North Atlantic glacial situation at around 18 ka BP (after Anderson and Borns, 1994). The British Ice Sheet has separated from the Fennoscandinavian Ice Sheet, though the North Sea remains dry due to low sea levels.

1.3.3: 17- 14.4-15.1 ka BP

Most authors agree that the large ice sheets surrounding the North Atlantic at this time were in a stage of rapid retreat (Figure 1.9). Lehman *et al.* (1991) state that the Fennoscandinavian ice sheet and the Barents Sea Ice Sheet were retreating from their LGM margins. Jones and Keigwin suggest that the rapid retreat of these ice sheet systems was accentuated by eustatic sea level rise, a theory which is backed by Veum *et al.* (1992), who propose that ice free conditions dominated the North Atlantic as warm water flow re-established itself. Both Maslin and Sarnthein agree that the production of NADW was at its lowest between 16.9 and *14.4-15.1* ka BP, though a sluggish conveyor system was still in operation. Hald and Aspeli (1997) identify a sharp warming phase at 15.6 ka BP, which is supported by the meltwater events described by Rasmussen, Lehman and Jones and Keigwin. At this time Atlantic water entered the Norwegian Sea and the continental shelf area of the Fennoscandinavian ice sheet was actively deglaciating. Maslin also identifies a period of rapid retreat of the Laurentide Ice Sheet. Hald and Aspeli suggest that modern temperatures were attained in the Norwegian Sea by *14.4-15.2* ka BP and Lehman and Keigwin (1992) propose that the polar front had retreated from its LGM position by this time.

*1.3.4: 14.4-15.1- 12.4-12.6*ka BP

From around 14.3 ka BP it is clear that warming was well underway, with Atlantic water entering the northern seas, and melting of all ice sheet systems surrounding the North Atlantic well established. By 14.1 ka BP the Polar Front had moved from 35° N to 55°N (Bard *et al.* 1987) and temperatures in the northern Norwegian and Barents Seas rose from 1° to 5.5°C (Hald & Aspeli, 1997). On the Rockall plateau, west of Ireland, SST had risen to modern temperatures, around 14°C by 12.2 ka BP (Bard *et al.* 1987). Sarnthein proposes that this is the time when the Holocene Gyre system became re-established in the northern North Atlantic.

1.3.5: c12.6 - c10.8 BP: the Younger Dryas

Most Authors agree that the Younger Dryas period began at around 12.4-12.6 ka BP (Broecker, 1989, 1990a, 1990b, 1992, Bard *et al.* 1987, Jansen & Veum, 1990, Lehman *et al.* 1991, Veum *et al.* 1992, Charles *et al.* 1996, Lehman & Keigwin, 1992, Bond *et al.* 1993, Sarnthein *et al.* 1994, 1995). Precise timing of the period is contentious as a radiocarbon decay plateau exists at around this time. The period is associated with a general cooling of the oceanic circulation in the North Atlantic, an increase in sea

Figure 1.9: The North Atlantic glacial situation at around 17-15 ka BP. Ice sheets surrounding the ocean basin are in rapid retreat, and sea ice begins to break up. NADW circulation becomes re-established. This period is dominated by huge iceberg rafting events (Heinrich events) accentuated by sea level rise (after Anderson and Borns, 1994).

Figure 1.10: The North Atlantic during the Younger Dryas cold period. Sea ice regrows to fill the Bay of Biscay, and continental ice sheets begin to regrow, or become rejuvenated. A land bridge still exists between the British Isles and continental Europe (after Anderson and Borns, 1994).

ice cover, and a breakdown or slowing of the thermohaline conveyor (Figure 1.10). Temperature began to rise once more at around 10.3 ka BP (Hald & Aspeli, 1997, Lehman *et al.* 1991, Bond *et al.* 1993, Sarnthein, 1994, 1995), until Holocene temperatures were attained between 10.3 and 10.2 ka BP. Sarnthein argues that the re-establishment of full thermohaline circulation would have released large amounts of CO_2 producing the plateaux on the ^{14}C decay curve.

1.4: Scotland

1.4.1: The Scottish Record

The record of the deglacial phase in Scotland is fragmentary and poorly understood, due mainly to a lack of dates. The history of the deglaciation is further complicated by an incomplete shoreline record, and complex isostatic response of the continental shelf to loading and unloading by the Scottish Ice Sheet. Other problems include the presence of a ^{14}C plateau in the Younger Dryas, and contamination of dates by so called 'old' carbon. The interpretation of landforms and sedimentological sequences is also frequently open to debate. Nonetheless a broad synthesis of current ideas will be presented here concerning the behaviour of the Scottish Ice Sheet between around 23.7 ka BP and the start of the Holocene at around 10.8 ka BP.

1.4.1a: Glacial Maximum

It is generally thought that maximum ice extent in Scotland was reached at around 23.7 ka BP (Boulton, *et al.* 1977, Jones and Keen, 1993, Lowe and Walker, 1997, and others). Most of Scotland was ice covered at this time (Figure 1.11), including areas such as Buchan (Hall and Sugden, 1987). Most marine evidence for this period has come from the continental shelf west of the Outer Hebrides, where ice extended nearly to the edge of the continental shelf, and from cores west and south west of Ireland. Other studies have shown that the Shetland Islands were extensively glaciated, and that most of the North Sea basin was ice covered. It is generally assumed that pack ice surrounded Britain's' shores until at least 16.2 ka BP.

The west coast of Scotland experienced maximum glaciation between around 23.7 ka BP and 22 ka BP (Austin and Kroon, 1994, Peacock and Harkness, 1990, Peacock and Long, 1994, Peacock *et al.* 1992, Peacock, 1984, 1992, 1997, Selby, 1989) with ice reaching the edge of the continental shelf west of the Outer Hebrides, and surrounding St. Kilda. Its maximum position is marked by a series of morainal banks

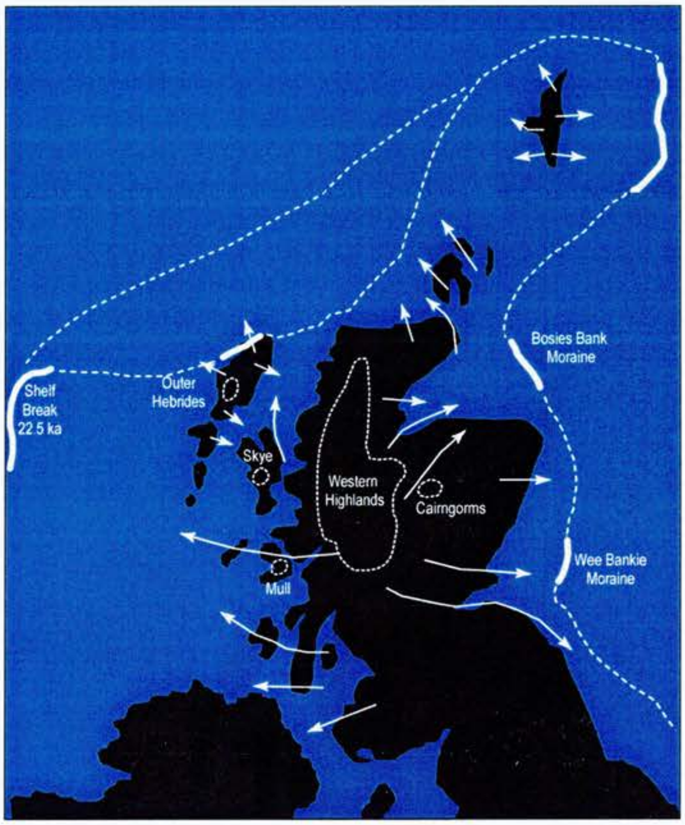


Figure 1.11: Summary of possible Scottish LGM glacial situation from various authors. Sea floor moraine limits (thick white line) described by Hall (1997), Peacock (1994) Peacock and Long (1995), Hall and Bent (1990). Proposed limits described by Hall and Bent (1990), and Lambeck (1993). Proposed spreading centres (small dotted lines) described by Sutherland and Gordon (1993), McCarrol et al. (1995), Ballantyne (1990, 1997), Ballantyne et al. (1987, 1997, 1998), Benn (1997) Ballantyne and Benn (1991). Ice flow directions (white arrows) proposed by Boulton et al. (1977) and Hall (1997).

(Figure 1.12) (Selby, 1989, Austin and Kroon, 1994), and has been AMS ^{14}C dated to 22, 480 ± 300 yr BP (beyond limit of Calib 4.3 curve, Stuiver *et al.* 1998).

Further north the St. Kilda plateau was ice covered, though the island itself was not extensively glaciated. There is evidence in the form of morainal banks 60– 80 km north of the island on the shelf break, to suggest that St. Kilda was glaciated earlier than the late Devensian (Stoker, 1988).

North of St. Kilda the grounded ice is thought to have extended north eastwards to the northern tip of the isle of Lewis (Peacock, 1984, Peacock *et al.* 1992, Sutherland and Walker, 1984, Von Weymarn, 1979). Peacock and Harkness (1990) identify a limit to the ice from the Scottish mainland at the position of a borehole off Stornoway. This implies a zone of confluence of two ice sheet systems, one from the mainland and one from the Outer Hebrides. It is possible to identify other areas of local spreading centres around the Scottish coastline, for example in Shetland.

There is very little information available for the far north of Scotland including the areas of Assynt, Caithness and Sutherland, though most reconstructions (Boulton, 1977, Boulton *et al.* 1985, 1991) suggest that ice flowed north of the present coastline, almost up to the continental shelf break. Lambeck (1993) proposes a limit just off the present north coast based on isostatic response, but this does not appear to fit with other evidence for the Orkney and Shetland Islands.

Several authors have assumed that the Shetland Islands were glaciated during the late Devensian, probably chiefly by local ice, but also by Scandinavian ice (Mykura, 1976, Flinn, 1978). Peacock (1994) describes sedimentological evidence that shows glaciation of the islands continued from the LGM up until the start of the Windermere Interstadial, with a proposed maximum limit some 60 km off the present eastern shores of the islands (Figure 1.13).

Peacock and Harkness (1990) describe the North Sea basin as having relatively high sea levels adjacent to glacierised areas, and relatively low sea levels elsewhere. Much of the North Sea was dry land, and the Scandinavian Ice Sheet extended across the Norwegian Channel, partially blocking a shallow sea that extended SW from east of Shetland and Orkney to near eastern Scotland, where it was in contact with the eastern margin of the Scottish Ice Sheet. Offshore of the major ice stream

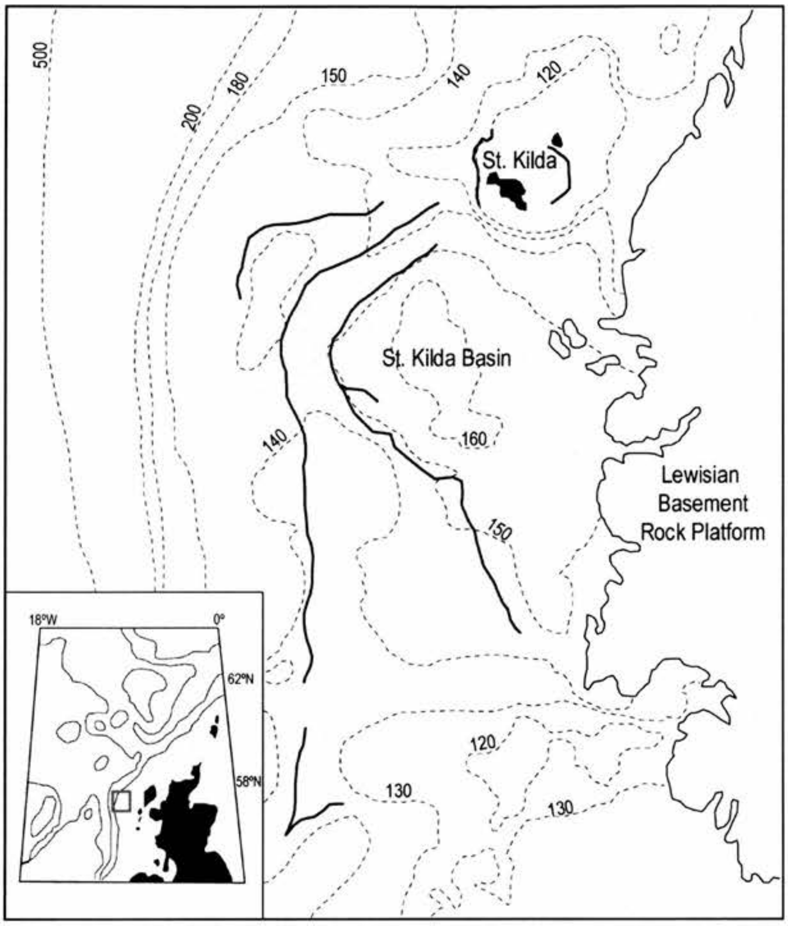


Figure 1.12: LGM morainal limits in the St. Kilda area. Continuous line: exposed limits of Lewisian Basement Rock Platform; Thick line: morainal bank complexes; Dotted line: bathymetry (after Austin & Kroon, 1994).

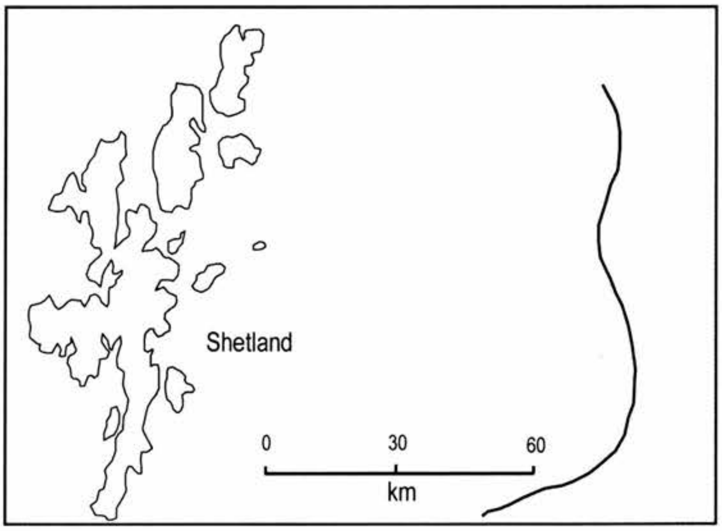


Figure 1.13: Eastward limit of Late Devensian glacigenic deposits off Shetland (after Peacock, 1994).

outlets there are moraine sequences identifying maximum Devensian limits of the Scottish Ice Sheet. Sedimentological evidence in the Moray Firth points to a two stage glaciation of the area during the Late Devensian, the first predating 23.7 ka BP, and the second between 17.3 and 16.2 ka BP (Peacock, 1997). Further south at the outlet of the Dee valley ice stream the Wee Bankie Moraine delimits the maximum extent of ice off Aberdeenshire (Brown, 1993). In the Firth of Forth authors (Sutherland, 1984, Cameron *et al.* 1987) have placed the maximum limit some 100 km to the east of the present coastline.

Inland most of Scotland supported a thick ice cap, with the main spreading centre over the western Grampians, and minor centres in Mull, the Outer Hebrides, the Cairngorms and the south east Grampians. Gordon and Sutherland (1993) argue that none of these were overridden by the main Highland ice mass. The thickness of the ice varied considerably, with some authors suggesting the presence of large numbers of nunataks in the west (McCarroll *et al.* 1995, Ballantyne *et al.* 1997, 1998), and thick ice cover in the Buchan area (Hall and Sugden, 1987). Though the Scottish Ice Sheet system can be compared with typical sub polar ice sheet systems present today, with large ice streams draining the major ice masses, which were centred on upland regions. Recent work points to an ice sheet somewhat thinner than that proposed by Boulton (1977) who assumed a maximum thickness of 1800 m of ice over the Highland region. Ice free summits have been proposed in the Western Highlands by Ballantyne, who suggests ice surface altitudes of 700 m on An Teallach, and identifies a proposed trimline extending as low as 450 m on the Trotternish Peninsula on Skye (Ballantyne and Sutherland, 1987, Ballantyne, 1990, 1997, 1998). Gordon and Sutherland suggest that the combination of evidence from ice-free areas along the northern fringes of the ice sheet, its 'modest' dimensions with distances between ice sheds to margins no greater than 200 km, and the smaller independent ice spreading centres on the Cairngorms, Mull and the Outer Hebrides, argue for a relatively thin ice sheet system (Gordon & Sutherland, 1993).

1.4.1b: Deglaciation

Most authors agree that rapid deglaciation in Scotland occurred at around 17.6 ka BP (Austin and Kroon, 1994, Peacock and Harkness, 1990, Peacock *et al.* 1992, Peacock and Long, 1994, Gray, 1995, and others). Koc *et al.* (1992) propose the presence of a seasonally ice free corridor off the western coast of Norway at this time.

Austin and Kroon (1994) suggest that deglaciation of the west coast began at 18.2 ka BP and was complete by 15.6 ka BP. They derive these dates from the dating and interpretation of sediment from boreholes in the St. Kilda Basin, which point to a period of glacier retreat, high meltwater and high sediment flux at the time. By 16.1 ka BP glaciers on the west coast probably terminated within sea lochs, and the major spreading centres, namely the western Grampian and the Hebridean ice caps, had separated (Peacock and Harkness, 1990). A shallow sea had opened up between the Outer Hebrides and the Scottish mainland, which was probably open to the Atlantic at the Butt of Lewis and in the south across the Malin shelf. Peacock *et al.* (1992) suggest that much of the ice in the St. Kilda basin area could have become unstable as a result of rising sea levels, and rapid calving could have led to the fast break up of the ice sheet here. They also suggest, along with Austin and Kroon (1994) that at least during the early part of the deglacial phase, isostatic readjustment might have kept pace with sea level rise, thus maintaining shallow seas, and leading to the rapid sea level falls observed later during deglaciation.

To the north, Shetland remained glaciated until the beginning of the Windermere Interstadial (Peacock and Long, 1994). Evidence from marine and lacustrine sediments indicate that deglaciation of the Shetland region was late compared to mainland Scotland, and that the Shetland Ice Cap was still within a few tens of kilometres of its maximum extent close to the beginning of the interstadial, with deglaciation proceeding rapidly thereafter. Problems are encountered with dating the material at this point in the record, as a 'radiocarbon plateau' exists (Stuiver *et al.* 1998). The marine reservoir effect is also a complicating factor, the extent of which is yet to be quantified. Peacock and Harkness (1990) suggest that there was a land bridge between the mainland and Orkney, and so too between Orkney and Shetland, with the eastern extent of this landmass being found close to the south west margin of the Norwegian channel. The sea to the east of this now spread as far south and west as the mouths of the Moray Firth and the Firth of Forth.

Polar water was present in the northern North Sea until at least 15.9 ka BP (Peacock and Harkness, 1990). A ^{14}C date from the St. Fergus silts in north east Scotland shows that ice had retreated to the present coastline by around 15.3 ka BP (18.3 cal BP, Hall and Jarvis, 1989, Brown, 1993, Peacock, 1997). Brown (1993) proposes that rapid retreat of the ice front was a result of accelerated calving due to sea level rise.

1.4.1c: Readvances

Fluctuations of the eastern parts of the ice sheet, and particularly periods of 'still-stand', can be seen in the Cairngorms (Sugden, 1975, 1980, Sissons, 1979a & b, Brazier *et al.* 1996, 1998, Everest & Kubik, *in prep*) and in the eastern Grampians (Brown, 1993). In the west a series of readvances, such as the proposed Wester Ross Readvance have been identified within proposed glacial maximum limits (Robinson & Ballantyne, 1979 Sissons & Dawson, 1981) (Figure 1.14). Central and eastern parts of Scotland have been argued to support evidence for early readvances of the Scottish Ice Sheet, such as the Aberdeen-Lammermuir Readvance (Sissons, 1967), though much of this evidence has been questioned in the literature (Gordon & Sutherland, eds, 1993, Gordon, ed, 1997). More often than not these limits are now interpreted as being sites sharing similar stratigraphies rather than distinct glacial geomorphology, and are not thought to represent regionally integrated ice sheet limits (Hall, in Gordon ed. 1997).

Further readvances may also be identified on the northeastern coast of Ireland, where the Armoy moraine in Antrim marks the position of the confluent Linnhe and Clyde ice streams. Clapperton (In Gordon, ed, 1997) suggests that this limit may be correlated with Sissons (1967) Aberdeen-Lammermuir Readvance, as well as limits across the Moray Firth at Lhanbryde and the lower Spey Valley (Brown, 1993). Clapperton argues that these limits are all older than 12.6 ka BP, indicated by ^{14}C data from basal bog sediments near Braemar, though they are younger than 15.4 ka BP, which has been cited as the minimum age for onshore recession of the coastal Aberdeenshire ice (Hall & Jarvis, 1989). Clapperton also argues that readvances in the Cairngorms identified by Brazier *et al.* (1996) are similar in size and stratigraphic position to the readvance at Lhanbryde.

The next readvance has only tentatively been dated at Wester Ross (Robinson & Ballantyne, 1979) to 16 – 14.3 ka BP, (age calibrated from a ^{14}C sample from Loch Droma, 'well within' the proposed Wester Ross limit) where an onshore limit marks the maximum position of a readvancing ice mass. This limit has been correlated with the Otter Ferry Stage limit in Loch Fyne (Sutherland, 1984, Clapperton, 1997). It has been argued that these events have been placed too early in the deglacial record, possibly occurring 1 ka earlier during H1, though no directly dated limits have been confidently assigned to this period in Scotland.

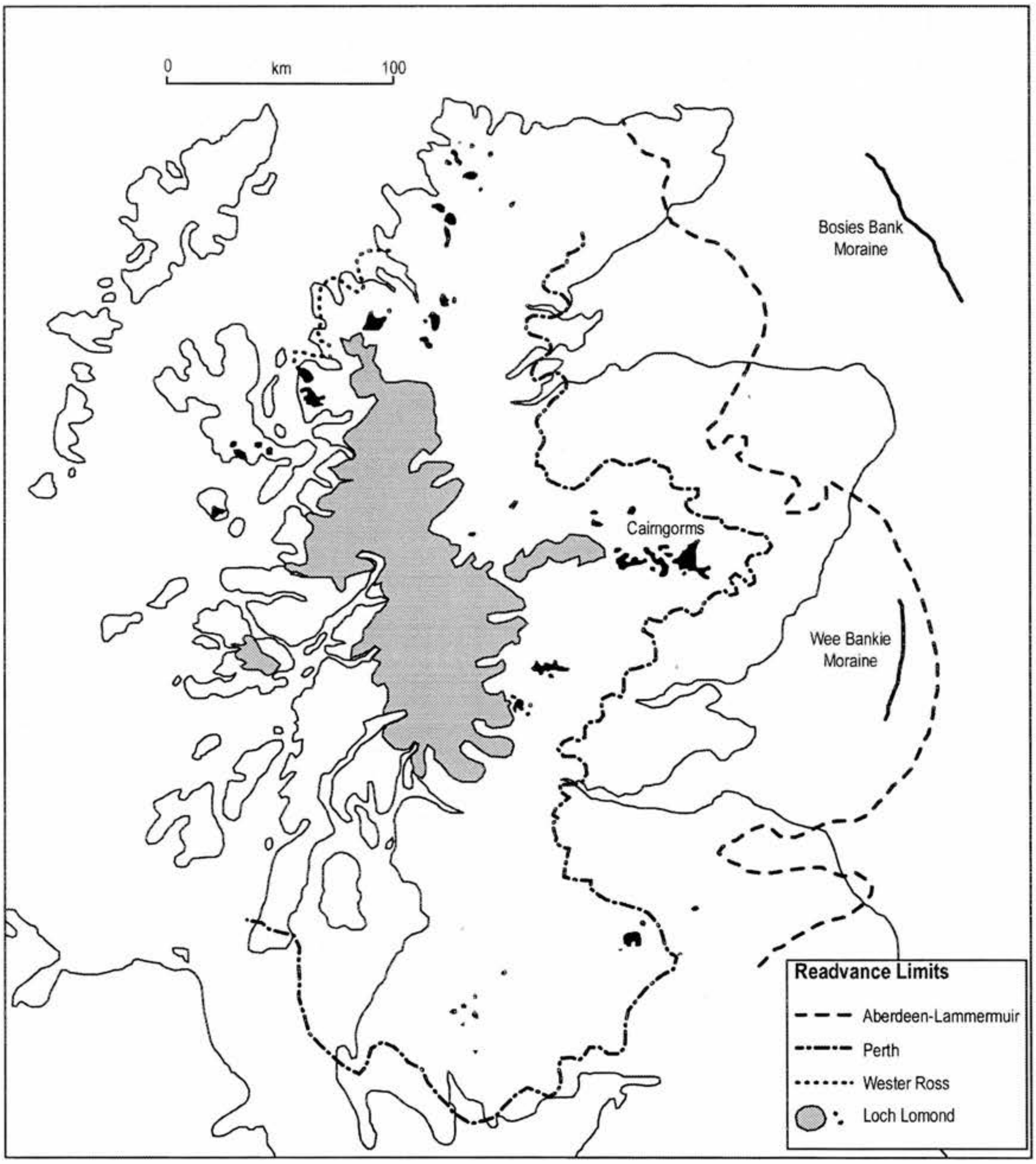


Figure 1.14: Proposed readvance limits in Scotland (Sissons, 1980, Robinson & Ballantyne, 1979).

1.4.1d: The Windermere Interstadial

Nearly all authors point to 16.1- 15.6 ka BP as being the point when full interstadial conditions were established in Scotland. By this time Scotland was surrounded by warm Atlantic water, and most glaciers had disappeared. Some authors suggest that icebergs were still present, particularly off the west coast, shown by traces of IRD within marine sediments (Peacock *et al.* 1992), though the source area for these was probably not the remnants of ice on the outer Hebrides or the Scottish mainland. There is still debate surrounding the survival of ice in the high corries of the Scottish mountains throughout the interstadial (Bennett & Glasser, 1991).

1.4.1e: The Loch Lomond Stadial

Several reconstructions of the extent of the Loch Lomond Stadial ice masses have been proposed. All agree that the western Grampians supported by far the largest ice sheet (Figure 1.14 and 1.15). With a major spreading centre in Rannoch Moor, it supplied ice that flowed under topographic constraint to the south, north and west, along with smaller centres as far north as Torridon (Bennett and Boulton, 1993, Fig. 6). Few glaciers reached sea level, the only proposed calving glaciers at this time were in Lochs Hourn, Nevis and Linnhe. The western ice sheet spread as far east as Achnasheen in the north (Benn, 1996, Thorpe, 1986, 1987), Glens Affric and Morriston, and The Great Glen (Bennett and Boulton, 1993, Sissons, 1979, Thorpe, 1986, 1987). The Outer Hebrides remained unglaciated, as did Shetland, though the Inner Hebrides, in particular Skye and Mull are said to have supported small ice caps (Ballantyne and Benn, 1994, Benn, 1997, Walker *et al.* 1988). Further north, Wester Ross and Assynt supported small corrie glaciers, though none developed into valley glaciers (Robinson, 1987, Sissons and Dawson, 1981, Lawson, 1986, Ballantyne, 1986).

In the east the situation was very different. Small ice caps have been proposed to exist on the Gaick and Monadhliath Plateaux, along with ice in Drumochter (Sissons, 1980) and the Loch Muick/ Glen Clova region (Sissons & Sutherland, 1976). Elsewhere only corrie glaciation was possible, and even here only the highest corries were ice filled. In the Cairngorms Sissons (1979) identifies seventeen corrie glaciers around the massif (Figure 1.19). Sugden (1980) proposes even fewer, with current work identifying even more limited glaciation of the Cairngorms during the Younger Dryas (Purves *et al.* 1999). Further south a corrie glacier existed in Lochnagar (Clapperton, 1986), and Cornish (1981) has identified 11 corrie glaciers in the Southern Uplands.

Figure 1.15: Scottish Loch Lomond Stadial limits proposed by Sissons (1974), Bennett & Boulton (1993), Thorp (1987) and modelled limits by Hubbard (1999). Few authors have attempted to reconstruct the whole extent of the Loch Lomond ice sheet in Scotland, and the differences apparent in these reconstructions illustrate why this has been the case. It is interesting to note that apart from the few exceptions visible in the Thorp reconstruction, only the modelled limits of Hubbard show significant ice free areas, and artefact of the 'ruling hypothesis' throughout the 1960's, 70's and 80's, that the Loch Lomond ice sheet was thicker than modern thinking suggests. None of the authors predict large scale glaciation in the east, with only Sissons (1974) proposing anything more than very minor corrie glaciation. Hubbard's model places small plateau ice caps predominantly in the Monadhliath, Gaick and Monadh Mor plateau south of Glen Einich.

1.4.2: Modelled Scottish Ice Sheet Data

Several authors have attempted to produce numerically derived models of Scottish glaciation (Boulton, 1977, 1985, 1991, Lambeck, 1993, Hubbard, 1999). Siegert (2001) states that the British Ice Sheet is one that is highly suitable to modelling, as there is a wealth of evidence, and good understanding of the processes involved in British glaciation. He argues that glacial geologic evidence such as striae and roches moutonnées, can provide important constraints to ice flow directions; centres of ice loading can be determined using uplift curves derived from raised beach studies (eg. Lambeck, 1993); ice thicknesses can be constrained in mountain areas by trimlines and summit erosion features (eg. Ballantyne, *et al.* 1997); finally Siegert states that subglacial conditions may be established by analysis of drumlins and meltwater systems (eg. Glasser and Sambrook-Smith, 1999). Typically however most models proposed thus far for British ice sheets are constrained by only a few points. For example Hubbards model (1999) utilises constraints from ice limits on Skye, Mull and the Cairngorms. All authors argue that more dated constraints are needed to more accurately model British ice sheet behaviour.

1.4.2a&b: Boulton 1977 and 1985

Boulton's first model (1977) is constrained by moraine limits in the south, and proposes that the British Ice Sheet was initiated in the uplands of Ireland, Scotland and Wales (Figure 1.16a). It is a flowline model, these being determined by field-based observation of the glacial geology. The model is viewed as a 'maximum' reconstruction (Siegert, 2001), with maximum thicknesses of 1800 m over large areas of upland Britain. It also proposes confluence with the Scandinavian Ice Sheet. A later model proposed by Boulton *et al.* (1985) is now viewed as a 'minimum' reconstruction of the British Ice Sheet at LGM (Figure 1.16b). In contrast to the earlier model, the eastern margin of the ice sheet terminates around 100 km off the coast, and is not confluent with ice from Scandinavia. The Irish Sea is simply covered by an ice shelf, rather than supporting an ice stream, as proposed in 1977. Upland areas support much thinner ice masses, up to 1000 m thick.

1.4.2c: Boulton, et al. 1991

A later iteration of Boulton's model attempted to reconstruct patterns of ice flow as a result of a thinner modelled ice mass. Similar to the earlier model, it is constrained by field based evidence, however ruling hypotheses of the time argued for thinner ice masses, and in particular the British Ice Sheet to be separate from the Fennoscandinavian Ice Sheet, with an ice shelf in the North Sea. Interestingly Boulton, *et al.* argue for the presence of nunataks at glacial maximum, particularly on the seaward ranges of the Western Grampians, where a significantly steeper ice sheet slope was modelled than for that in the east.

1.4.2d: Lambeck 1993

Rather than utilising glacial geology to constrain ice limits and thickness, Lambeck (1993) employs a range of isostatic data from raised beaches around the British coastline. Interestingly the model predicts little or no ice in the North Sea, and a relatively thin British Ice Sheet at LGM, of no more than 1500 m thickness (Figure 1.16c). In this respect it supports a 'minimum' ice sheet similar to that proposed by Boulton *et al.* (1985).

1.4.2e: Hubbard 1999, Loch Lomond Stadial model

Recent model data describing the extent and dynamics of the Loch Lomond Stadial ice sheet in Scotland have been provided by Hubbard (1999), who uses a 1 km resolution 3D flowline model, driven by the GRIP temperature series. The model shows that under these conditions, adjusted for local variation in temperature, the Loch Lomond Stadial ice sheet reaches similar limits to those mapped (Ballantyne, 1989, Bennett, 1991, Sissons, 1974, Sutherland, 1984, Thorp, 1986, and Green, 1992) after 550 years (Figure 1.15d). After this time the modelled ice sheet continues to grow, in contrast to the situation represented by the geomorphology, however Hubbard argues that the southward migration of the Polar Front, leading to increased aridity in Scotland, led to the maintenance of a steady state system, which he effectively models by employing a reduction in precipitation to the model after this period. The model is interesting as it matches limits described by field reconstruction very effectively in the main western ice sheet, moreover it does not produce any significant ice growth in the major upland areas of the east of Scotland, such as the Cairngorms, Lochnagar or the Gaick.

1.5: Factors Affecting The Scottish Ice Sheet System

There are three major influences on the Scottish Ice Sheet throughout the deglacial phase arising from changes in North Atlantic circulation, namely SST, the state of the FIS and sea ice cover. Other significant influences on the behaviour of the ice sheet are atmospheric circulation and sea levels.

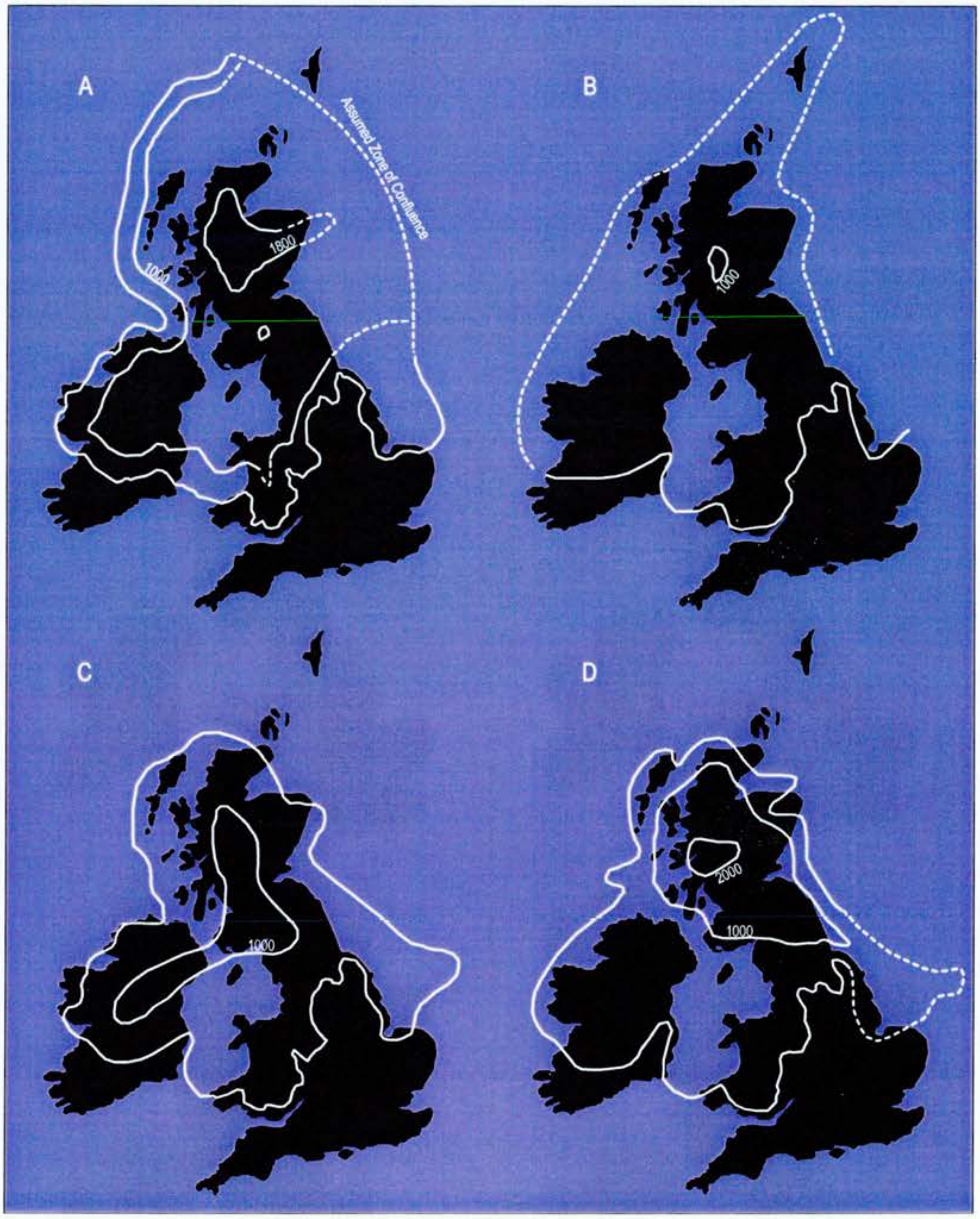


Figure 1.16: Modelled ice maxima at LGM. A: Boulton (1977); B: Boulton (1985); C: Boulton (1991); D: Lambeck (1993).

1.5.1: Oceanic Influences

1.5.1a: Sea Surface Temperatures

The temperature of the sea surface off Scotland is important in two respects. In the first instance it helps determine the type and volume of precipitation over Scotland, and secondly it determines the character and track of weather systems crossing the country. The two are obviously interlinked. The movement of the Polar Front in the North Atlantic (Figure 1.17) effectively determines SST for the region, though the behaviour of the FIS and the proposed drainage of the temporary Baltic meltwater lake will also greatly affect temperatures in Scotland. The history of wandering of the front gives some insight into the forcing that affected the Scottish Ice sheet system during the deglacial phase. Warming in SST from 18.3 to *14.4-15.2* ka BP correlates with a general decline of land based ice, with an almost complete disappearance of ice by 13.5 ka BP, and is in line with the 'modern' temperatures of the Sea Surface west of Ireland (Hald & Aspeli, 1997). Cooling of the sea surface from around 12.4-12.6 ka correlates with regrowth of ice in upland regions of Scotland at the start of the Younger Dryas. By most reasonable estimates there was a lag of between 500 and 1000 years between the reduction in SST and the time of regrowth of ice, though the ending of the Younger Dryas in the marine record fits very closely with the final disappearance of ice from Scotland.

1.5.1b: Sea Ice Cover

There is some disagreement in the literature over the amount and timing of sea ice cover during the deglacial phase. Earlier research assumes total ice cover in the Norwegian, Icelandic, Barents and Kara Seas until around 16.8 ka BP. This would obviously have huge consequences for the climate of Scotland and the extent of ice cover. Atmospheric temperatures are severely depressed, producing snow precipitation, as opposed to rain, though the volume of precipitation is much reduced. Studies have shown that large areas of ice cover create zones of moisture starvation downstream of the path of weather systems (Hubbard, 1999). The amelioration of temperature after LGM up to 16.2 ka BP was marked by the retreat of land based and ice shelf ice both in Europe and Scotland. With seasonally ice free seas other factors, such as increased snow precipitation must have been at work to prevent complete ice wastage over a very short space of time. Depressed

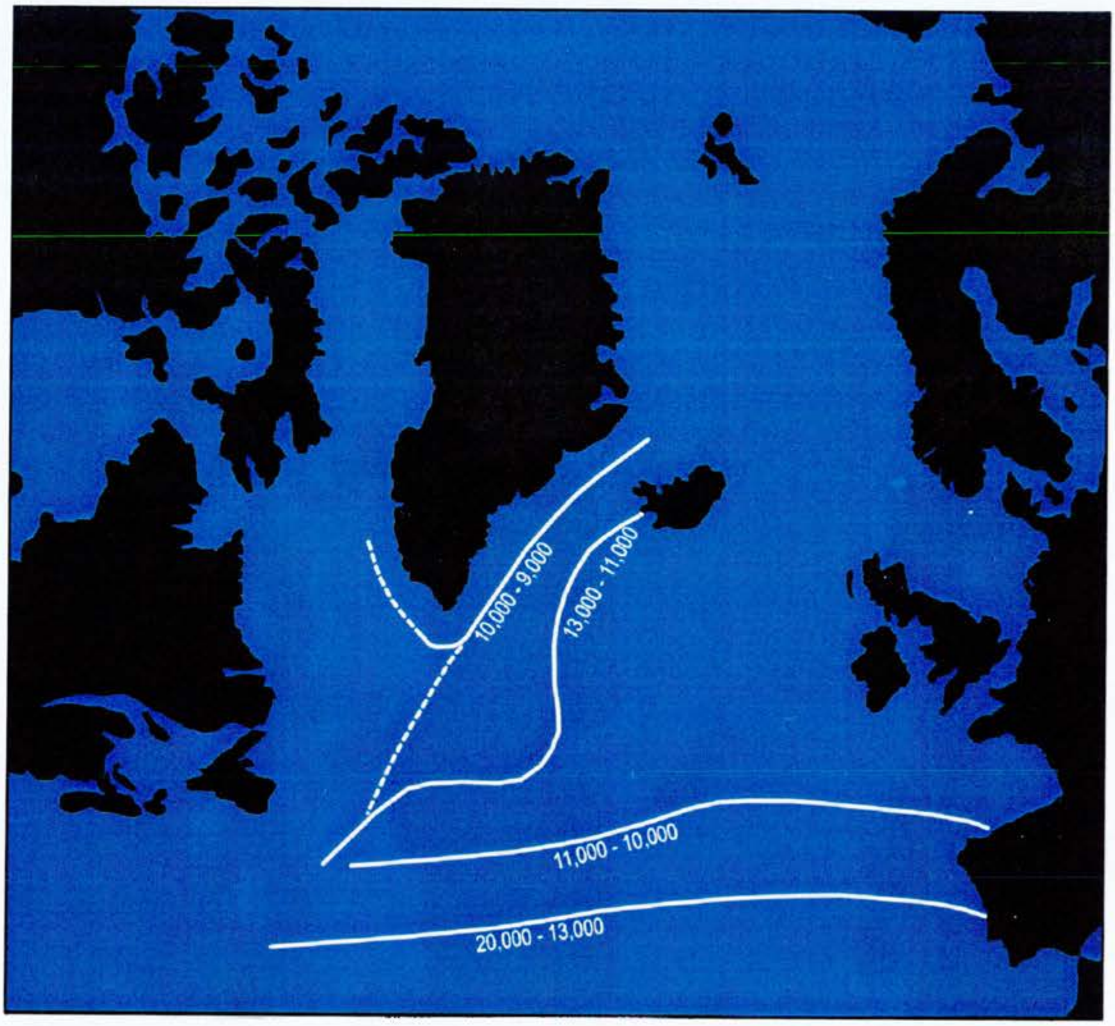


Figure 1.17: Location of the North Atlantic Polar Front throughout the LGIT (after Ruddiman and Macintyre, 1981).

SST and the influence of the diversion of North Atlantic surface flow, in combination with large amounts of icebergs in the area as a result of the collapse of ice sheet margins around the North Atlantic, were also dominant forces in the slowing of deglaciation.

1.5.1c: The Collapse Of The Fennoscandinavian Ice Sheet

Recent work points to the influence of the wastage of the FIS in the cooling regime of the North Atlantic (Sarnthein, 1994, 1995, Lehman *et al.* 1991). Meltwater from the FIS drains both into the Norwegian Sea and when open, the North Sea. Such an influx of cold water in the first instance cools the surroundings, and also it is argued it can extend further westward into the Atlantic and thus block the flow of warmer water from the south west, thus continuing the glacial conditions prevailing during the LGM. The collapse of the FIS has been cited as a cause of the Younger Dryas, as has the drainage of the meltwater lake in the Baltic region, by reducing the salinity and temperatures of the surface waters in the area, thus halting convection. It is plausible that a more complete record of the wastage of the FIS may provide an analogous chronology to the still stands and readvances proposed for the Scottish Ice Sheet system.

1.5.2: Atmospheric Influences

The role of atmospheric fluctuation in changing environmental conditions in Scotland during deglaciation is beyond question, however it is harder to quantify than the activity of the oceanic systems. Links can be made between Greenland ice core records and local vegetational and faunal records, though precise conclusions cannot be drawn due to the nature of such records. Estimations of temperature and precipitation change during the deglacial phase vary depending on the source locales for such records, though the movement of the atmospheric Polar Front northward throughout the phase profoundly affected the climate of Scotland.

What is universally accepted from the Scottish records is that at around 13 and 10 ^{14}C ka the rapid northward movement of both the oceanic and atmospheric Polar Fronts was accompanied by a significant temperature rise. At 13 ^{14}C ka a reconstructed temperature rise of 5 - 6° occurred in England, Wales and southern Scotland, resulting in Interstadial annual mean temperatures being only slightly cooler than those of today (Atkinson, *et al.* 1987). The increased maritime influence provided by the clearing of pack ice from coastal areas led to an accompanying increase in precipitation. A similar signal is present for the end of the Younger Dryas period, however the onset and duration of the Younger Dryas were characterised by intensely cold and dry atmospheric conditions in Scotland. Winter temperatures may have been as low as -20 to -30°C, and the period was characterised by increasing aridity, shown by pollen, lacustrine, coleopteran and diatomaceous evidence (Lowe and Walker, 1997).

1.5.3: Sea Levels

Both on and offshore, sea level variation profoundly affected the Scottish Ice Sheet. Offshore in the west the break up of ice on the continental shelf and in the Minch took place between 22 and 13 ^{14}C ka BP. This was accompanied by the invasion of the areas by sea, increasing water depths gradually from 40 m to modern depths of around 150 m on the shelf (Austin and Kroon, 1996, Peacock *et al.* 1992). The North Sea saw repeated marine invasions in the Viking Bank area (Peacock, 1995), though water depths rose to within 80 m of modern depths by 12 ^{14}C ka BP.

Onshore the earliest unequivocal evidence of shorelines associated with ice sheet decay are to be found between Stonehaven and the Firth of Forth, though later shorelines have been identified near Stirling and Perth (Sissons *et al.* 1966). Merritt *et al.* (1995) suggest a pattern of rapid retreat phases in the Moray Firth, in response to rising sea levels. West coast shorelines have been identified at many different locations; on Islay and Jura associated with a late Devensian limit across the islands, and possibly associated with the highest limit in East Fife (Dawson, 1982, cited in Smith, in Gordon, ed, 1997). Shorelines in Wester Ross have been associated with the Wester Ross Readvance by Sissons and Dawson (1979), and have been tentatively correlated with the main Perth Shoreline. These shorelines have also been associated with falls in relative sea level at Otter Ferry, Glendaruei, Loch Long, Loch Morar and Arisaig, each having falls of similar magnitude (Sutherland, 1981). Smith (1997) suggests that the correlation of these limits may reflect an ice sheet readvance, or simply the stabilisation of the retreating ice sheet as it became pinned at valley mouths.

1.6: The Cairngorm Mountains

1.6.1: Introduction

Controversy has surrounded the decline of the last ice sheet systems in eastern Scotland at the end of the last glacial period, with a main area of debate being the extent of glaciers in the major valleys and corries of the high Cairngorms during the Loch Lomond Stadial, or Younger Dryas. Some authors have argued for relatively extensive valley glaciers in Glen Geusachan and Upper Glen Dee during the Loch Lomond, whilst others have maintained that glaciation was confined to only the highest corries with favourable orientations. New evidence points to a more complex deglaciation sequence, with a major stillstand at around 15 ka BP, followed by ice wastage and the eventual disappearance of glaciers from all areas. This was then followed by the regrowth of ice in high corries during the Loch Lomond, with no major valley glaciers. The interpretation of important landform sequences, notably 'hummocky moraine', as indicators of different periods of glaciation, will need to be re-examined in the light of this evidence.

The Cairngorm mountains form the highest upland area in Britain, with several summits over 1200m (Figure 1.18). Today the climate is sub Arctic, with mean annual temperatures on the summit plateaux close to zero. Precipitation levels are lower than the western Highlands, which act as a barrier to the predominantly south-westerly precipitation bearing winds.

The plateau is dissected by several deep glaciated valleys, which reflect the passage of Scottish ice sheets which overrode the area during the Late Devensian and during earlier periods of glaciation.

1.6.2: Previous work

1.6.2a: Hinxman and Anderson, 1915

The first summary of the deglacial history of the Cairngorms was set out by Hinxman and Anderson (1915), as part of the first comprehensive geological survey of the region. They identified four distinct phases in the most recent glacial sequence:

1. Period of maximum glaciation; where the whole area of the Cairngorms and surrounding were covered by an eastward moving ice sheet, sourced from an ice shed to the west.

2. Period of large, confluent glaciers; large ice masses flowing through the area, sourced from the western ice centres, which were contemporaneous with the development of a radial spreading centre, based on the Cairngorm plateau. Hinxman and Anderson argued that these ice masses coalesced,

and flowed towards the north-east, north and north, north west, across the Monadhliath plateau.

3. Period of separation of the different ice systems, into independent valley glaciers; the Spey valley glacier continued to be fed from the west during its retreat, resulting in a multitude of retreat and depositional forms in and around the Spey and Glen More.

4. Period of high level corrie glaciation, with no ice occupation of major valley systems.

A key conclusion drawn by Hinxman and Anderson is that the Cairngorm Massif became an independent ice centre very early during the deglacial phase. Their evidence for this lies in the lack of erratics of non-local origin in valleys such as Glen Geusachan, the local ice having removed any schist erratics following the early phase of ice cover.

Hinxman and Anderson identify four sets of landforms that are key in the interpretation of the deglacial landscape of the Cairngorms.

1. Drainage channels running along contours, or transverse to the direction of natural drainage.

2. Lateral moraines and high level, ice-marginal terraces.

3. Valley floors covered with thick sequences of sands and gravels, following glacial retreat.

4. Ice stagnation forms, including kettle holes and collapse structures as a result of downwasting of dead ice blocks. (p. 60)

They also identify a complex interaction between local Cairngorm ice, and the larger regional ice sheet system. They attribute channels along the northern flanks of the Cairngorms to a downwasting ice mass in Glen More, and identify areas showing evidence of palaeo lakes in Glen Einich and the Lairig Ghru, dammed to the north by moraines where the Glen More ice pushed into those valleys.

They describe the character of retreat as being punctuated by "many pauses" (p. 61), indicated by the many parallel series of channels and terraces, taken to indicate the lateral position of the Glen More glacier. The highest of these terraces they propose runs from Castle Hill to Màm Suim, at an elevation of around 700 m. They argue that earliest retreat from the confluence of the Glen More glacier with that of the Nethy and other valley glaciers surrounding the basin is indicated by series of moraines, particularly in the Pass of Ryvoan. Further retreat is shown by extensive kettle holes, outwash gravels and finally "in the hollow which now holds the waters of Loch Morlich and represents the cast of a last remnant of the stagnant glacier" (p. 65).

The final glacial phase identified by Hinxman and Anderson is that of occupation of the highest corries by glacial ice. Even without local regional or hemispheric records of climate change, they still were able to state that the appearance of landforms resulting from these high corrie glaciers indicated a *regrowth* of ice, following the main ice sheet collapse. They do however cite the persistence of nevé in Coire an Lochan during the hot summer of 1893, acknowledging the possibility of long term ice survival in critical locations.

1.6.2b: Charlesworth, 1956

Charlesworth also recognised the relationship between the main Scottish Ice Sheet in the Spey, and the local Cairngorm ice, and suggested three broad phases of glacial retreat in the region, punctuated by shorter periods of rapid retreat.

1. A period of recession of the "vast" (p. 813) Spey Glacier from its coastal limits, depositing series of lateral moraines and lake terraces in the Spey Valley as a result of the impounding of meltwater from retreating ice lobes from the surrounding hillslopes. Charlesworth seems to suggest that ice on plateaux such as the Gaick, Monadhliath and Cairngorm was confluent with Spey ice, though never overridden by it, contrary to later theories (Boulton, 1977). The Kincardine Hills were overridden, though as the ice retreated in the Spey, their summits were revealed, from east to west, and the margin of the Spey Glacier occupied the Abernethy embayment to the north.

2. A decoupling phase followed in Glen More, characterised by the parting of the Cairngorm valley glaciers from the Glen More lobe, and the retreat of the Cairngorm glaciers into their glens. Charlesworth argues that the Glen More lobe advanced into the northerly parts of these glens, as seen by the moraines in the mouths of the Lairig Ghru and Glen Einich. He also argues that meltwater channels such as the Chalamain Gap were formed beneath the Glen More ice, and that the northern flanks of the Cairngorms were dotted with ice dammed lakes at the mouths of the major glens.

3. The final phase of main ice sheet deglaciation could be seen to the south west of the Cairngorms in the Spey Valley with series of moraines on the valley sides, back to the point of separation of the Truim glacier from the Spey. In the Cairngorms Charlesworth recognises a phase of retreat back to corrie glacial limits, with the Lairig Ghru glacier receding into the corries of Braeriach and Ben Macdui, and the Einich glacier receding and splitting into three separate lobes; a lobe draining the western corries of Braeriach; a series of corrie glaciers around Loch Einich; and a small glacier draining from the Moine Mhor plateau.

1.6.2c: Sugden and Sissons, 1968-81

Sugden believed that the overall summit geomorphology in the Cairngorms was created by overriding ice, though argued that this may not have occurred during the Late Devensian, and may indeed be the relic of much earlier glacial periods (1970). His interpretation of many of the features in the glens was that of an ice mass, downwasting in situ, with much of the geomorphology, in particular the 'hummocky moraine', being formed by topographically constrained meltwater at the end of the Late Devensian. He agreed that strong evidence existed for the interaction of the Scottish Ice Sheet in the Glen More basin, and Cairngorm ice in the northern glens.

He proposed that evidence existed in the high corries for a later stage of glaciation. Sugden (1970) applied a Scandinavian downwasting deglaciation scenario to the Cairngorms, and argued that the Loch Lomond Readvance in the Cairngorms was problematical. He identified two possibilities:

1. The area could have supported a significant ice sheet until some years after the Loch Lomond Stadial.

2. The forms in the area represented an earlier glacial event than the Loch Lomond Readvance, and the Loch Lomond limits were restricted to the corries (Figure 1.18).

When ^{14}C data from Loch Etteridge was published by Sissons and Walker (1974), this first possibility became untenable. Loch Etteridge is a small kettle hole, bounded by esker and kame systems (Walker, 1975) 25 km south west of the mouth of Glen Einich, on the south eastern side of the Spey Valley, near Newtonmore. The site was cored to enable a palynological reconstruction of the region, constrained by radiocarbon data. The conclusion of this study was that the site represented a sequence of lateglacial age, which had not been overridden by ice since 13,151 ± 390 ^{14}C BP.

The evidence used to describe the Loch Lomond Stadial advance was based on the interpretation of certain features in several of the valleys as being fluvioglacially derived. 'Hummocky till' was explained by the stagnation of ice in valley floors and much material was deposited by fluvioglacial activity beneath stagnant valley glaciers.

Sissons criticised this work, first in 1973, then after a reply from Sugden in 1975, with a further reassessment in 1979, and argued against Sugdens' reiteration (Clapperton *et al.*. 1975) of the existence of only small corrie glaciers in the Cairngorms during the Loch Lomond Readvance. He proposed the presence of seventeen glaciers of differing sizes in the Cairngorms (Figure 1.19), and argued that Sugdens interpretation of many glacial features as fluvioglacially derived was flawed.

Sissons argued that two distinct landform facies were to be found all over the Cairngorms- the first was associated with the decay of the Late Devensian Scottish Ice Sheet, and the second 'hummocky moraine' formed by the areal stagnation of the Loch Lomond Stadial readvance glaciers (1979). He identified large areas of this key landform assemblage in Garbh Coire, Glen Eidart and Glen Geusachan, where he proposed the existence of three significant Loch Lomond Stadial glaciers.

Sugden countered Sissons criticisms (Sugden, 1980). The hummocky moraine remained the crux of the argument for the delimiting of the Loch Lomond Advance, and Sugden, along with others (Clapperton *et al.* 1975) argued that such moraine in other parts of the surrounding area conformed to the decay of the remnants of the last major ice sheet in the area. He argued further that the limits proposed by Sissons for the glaciers of the Loch Lomond Readvance remained highly questionable. The debate has remained unresolved.

Young published a description of the collapse of the Glen More lobe of the Spey Valley glacier (Young, 1974). He described the series of meltwater channels that run along the northern flanks of the Cairngorms, leading into the Pass of Ryvoan, and attributed them to the wastage and collapse of a thinning Glen More lobe at the end of the LGM (Figure 1.20a). He identifies a period of gradual thinning, followed by widespread ice wastage and stagnation, concluding with a stage of collapse, leading to the creation of kettle holes, and outwash deposits.

1.6.2d: Recent work

More recent reinterpretation of the evidence, including work by Bennett and Glasser in 1991, proposed that the hummocky moraine is a result of ice-marginal deposition, which marks the successive stages in the retreat of active ice rather than the limit of stagnant ice as proposed by Sissons and Sugden. They go on to propose a regime for the glaciation of Glen Geusachan that features the survival of ice in the valley during the Windermere Interstadial, and its rejuvenation during the Loch Lomond Stadial (Figure 1.20b). In doing so they suggest that neither Sissons' theory of complete ice wastage after the last main glaciation, followed by regrowth in the valleys of the Cairngorms, nor Sugdens theory of a stillstand prior to the Loch Lomond Stadial, in the decay of a local plateau ice cap are correct.

The most recent review of glacial activity in the Cairngorms was carried out by Brazier, Kirkbride and Gordon in 1998, who reconstructed the deglaciation of the northern Cairngorms from east to west, at a time when ice from the main Scottish Ice Sheet was thinning and retreating up the Spey Valley, and in the Glen More basin. They confirmed the existence of a series of ice dammed lakes in the Lairig Ghru and Glen Einich, dammed to the north by ice in Glen More (Figure 1.20c). These lakes formed in a temporal sequence from east to west, as the Cairngorm ice thinned and detached from the Scottish Ice Sheet, forming an independent ice cap in the western Cairngorms. They describe the sedimentary sequences found in lacustrine

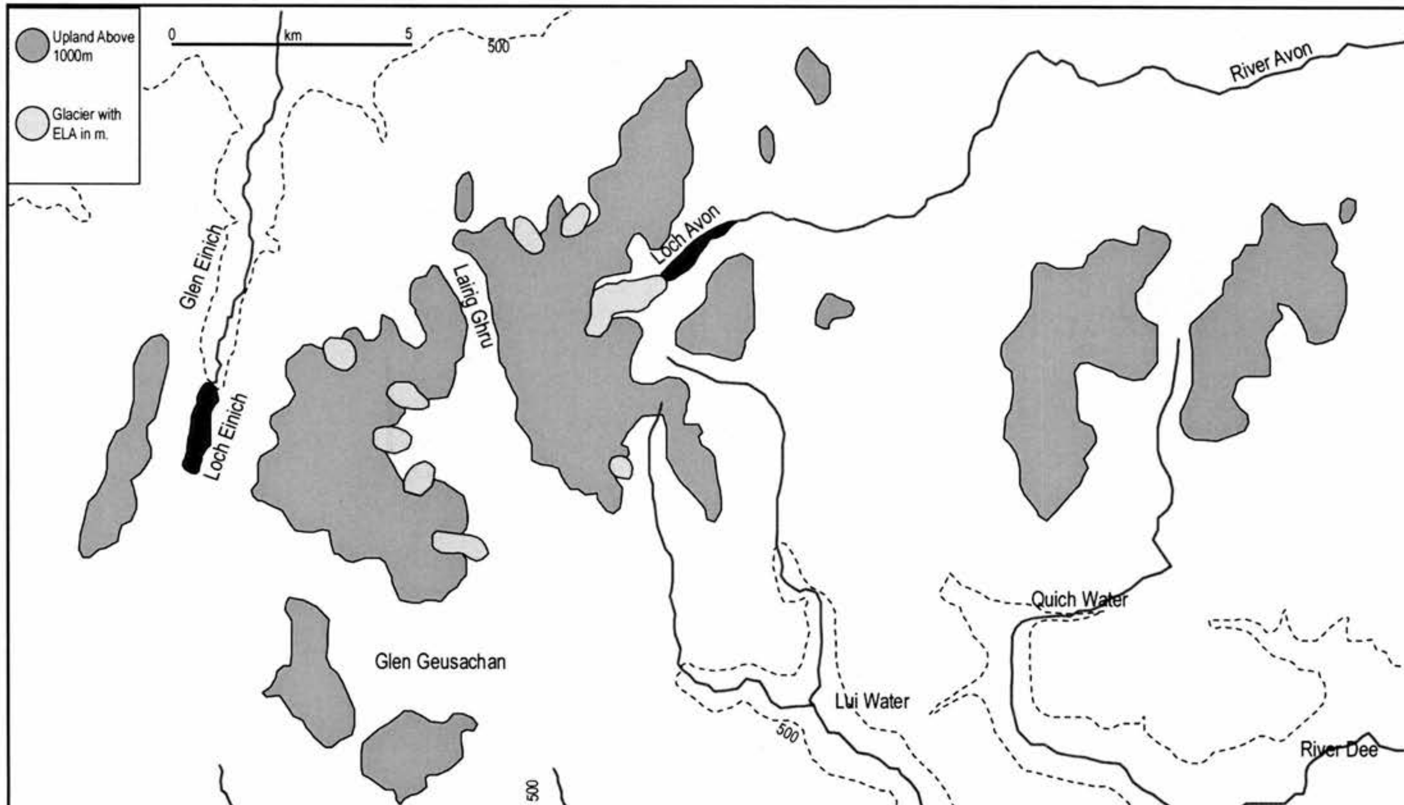


Figure 1.18: Loch Lomond Stadial glacial situation in the Cairngorms according to Sugden (1970).

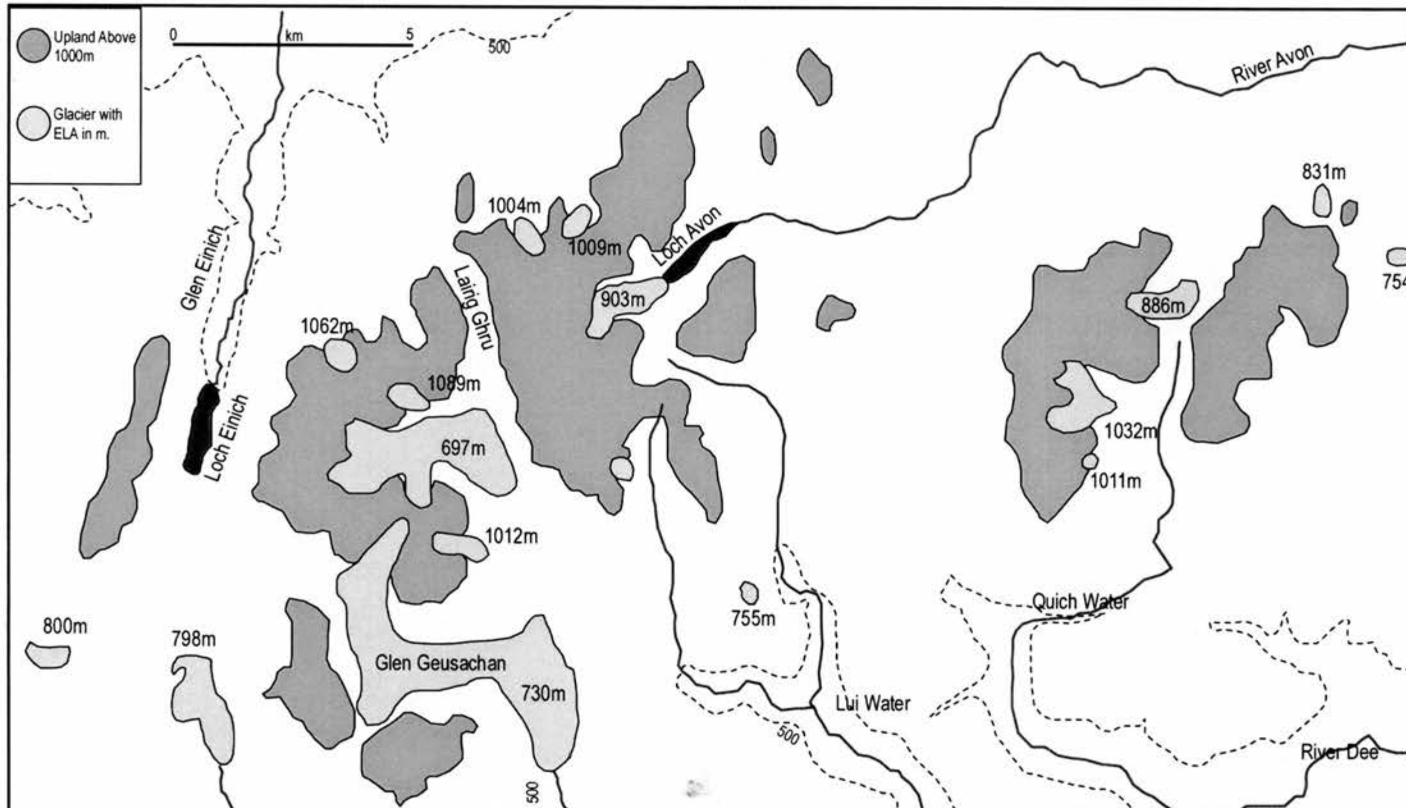


Figure 1.19: Loch Lomond Stadial glacial situation in the Cairngorms according to Sissons (1979).

deposits in these glens as being representative of still stands in the deglaciation of the Cairngorm ice cap. The stillstands were of the order of a few centuries (Brazier *et al.* 1998; p. 308), which make them potentially significant, in terms of the response of the Cairngorm ice sheet systems to more regional climate change.

In summary Brazier *et al.* (1996, 1998) describe a multi-stage pattern of retreat in the Cairngorms:

1. The initial thinning of the Scottish Ice Sheet revealed summits principally in the east of the massif. Smaller independent thin ice caps may have developed on the plateaux. Strath Nethy began to become deglaciated at this time, though ice in the Lairig Ghru was still confluent with the Scottish Ice Sheet ice in Glen More. The Glen More lobe became separated from the Nethy lobe by thinning of the Spey Ice to reveal the ridge to the south of the Ryvoan Pass.
2. an ice dammed lake. Sedimentary evidence in the Lairig Ghru indicates this lake was in existence for some time, perhaps of the order of centuries. Ice fronts must have remained relatively stable for this period to allow the persistence of the lake. Glen Einich remained full of ice draining northward from the western Cairngorm plateau, and the Feshie Hills emerged as a nunatak as the Spey Ice thinned further. Brazier, *et al.* argue that the Glen More ice remained active, discharging meltwater through Ryvoan, and creating the final sequences of glaciofluvial material in Abernethy.

   Separation of the Glen Einich glacier from the Glen More lobe then occurred, forming a further ice dammed lake. By this time the Glen More ice had retreated sufficiently from its margin at the mouth of the Lairig Ghru to allow the Lairig lake to drain eastward along the margin of the Glen More ice. This lake also persisted for a number of centuries, indicating relatively stable ice margin positions in Glen More. The ice in Glen Einich was supplied from the south west, over the Moine Mhor, which also fed the Glen Geusachan glacier at this time. Brazier *et al.* point out that both the Garbh Coire and Glen Geusachan glaciers were almost certainly contemporaneous with the Glen Einich glacier, despite being mapped as Loch Lomond Stadial features by Sissons (1979).

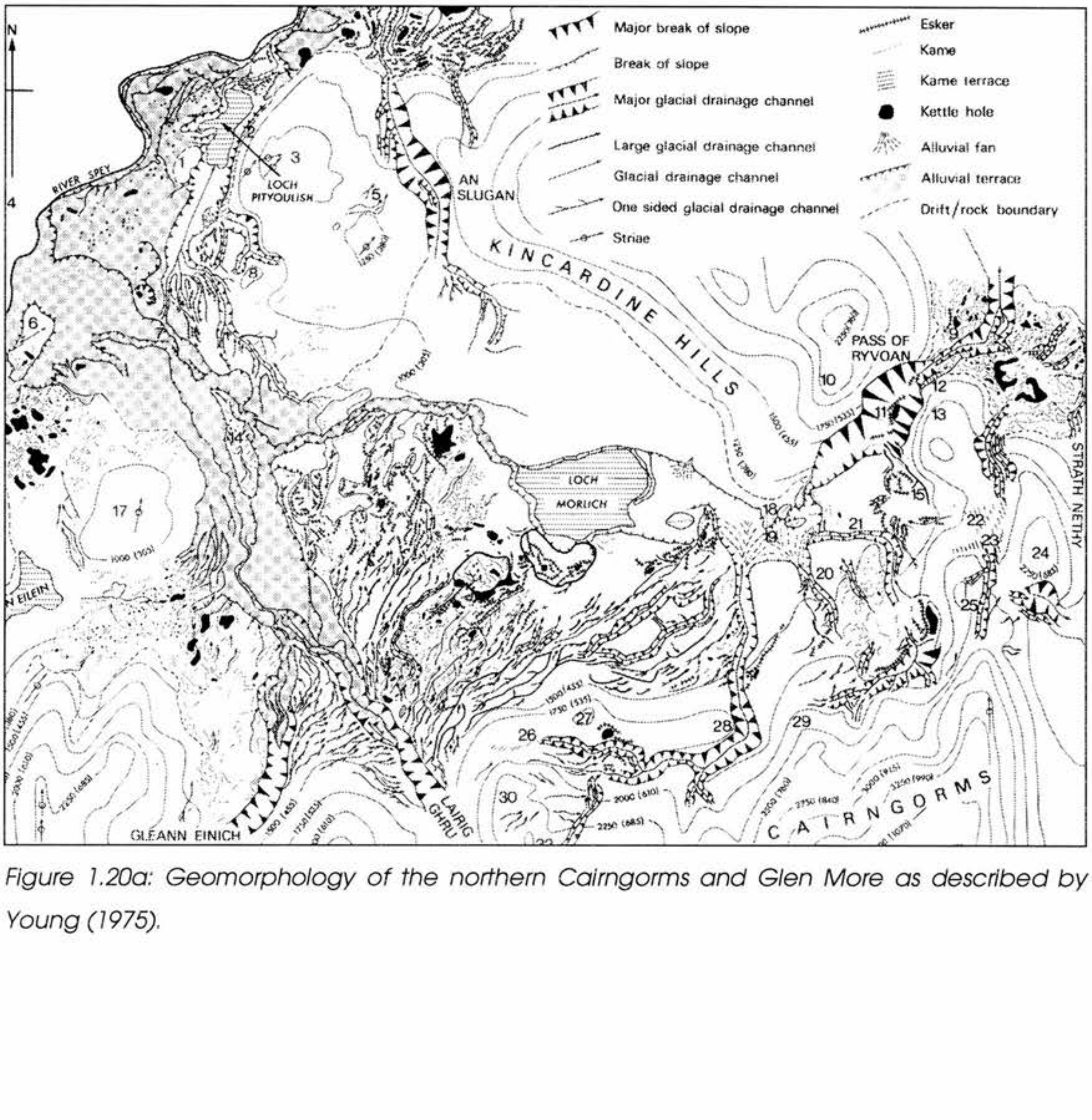


Figure 1.20a: Geomorphology of the northern Cairngorms and Glen More as described by Young (1975).

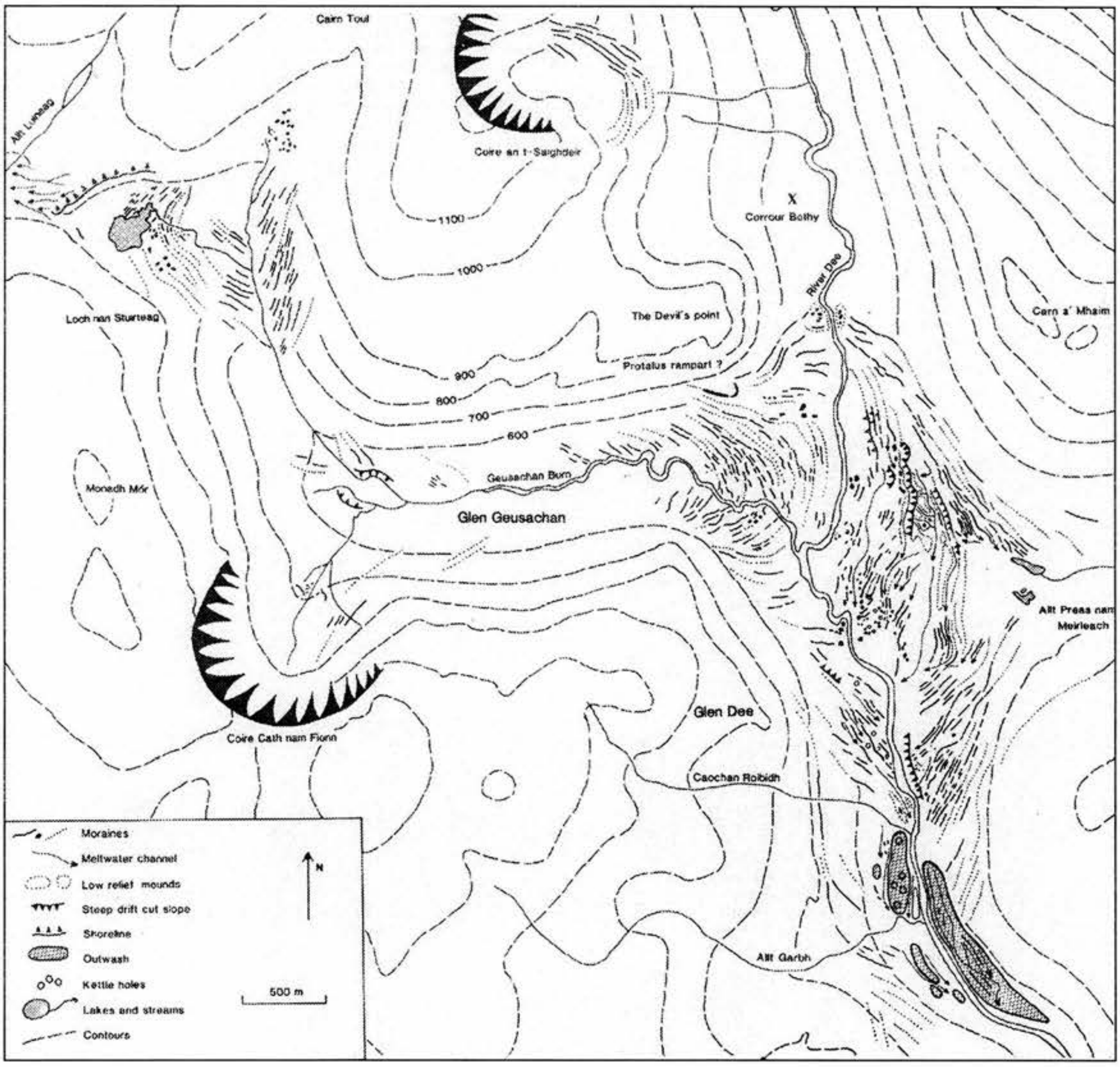


Figure 1.20b: Geomorphology of Glen Geusachan as described by Bennett & Glasser (1991)

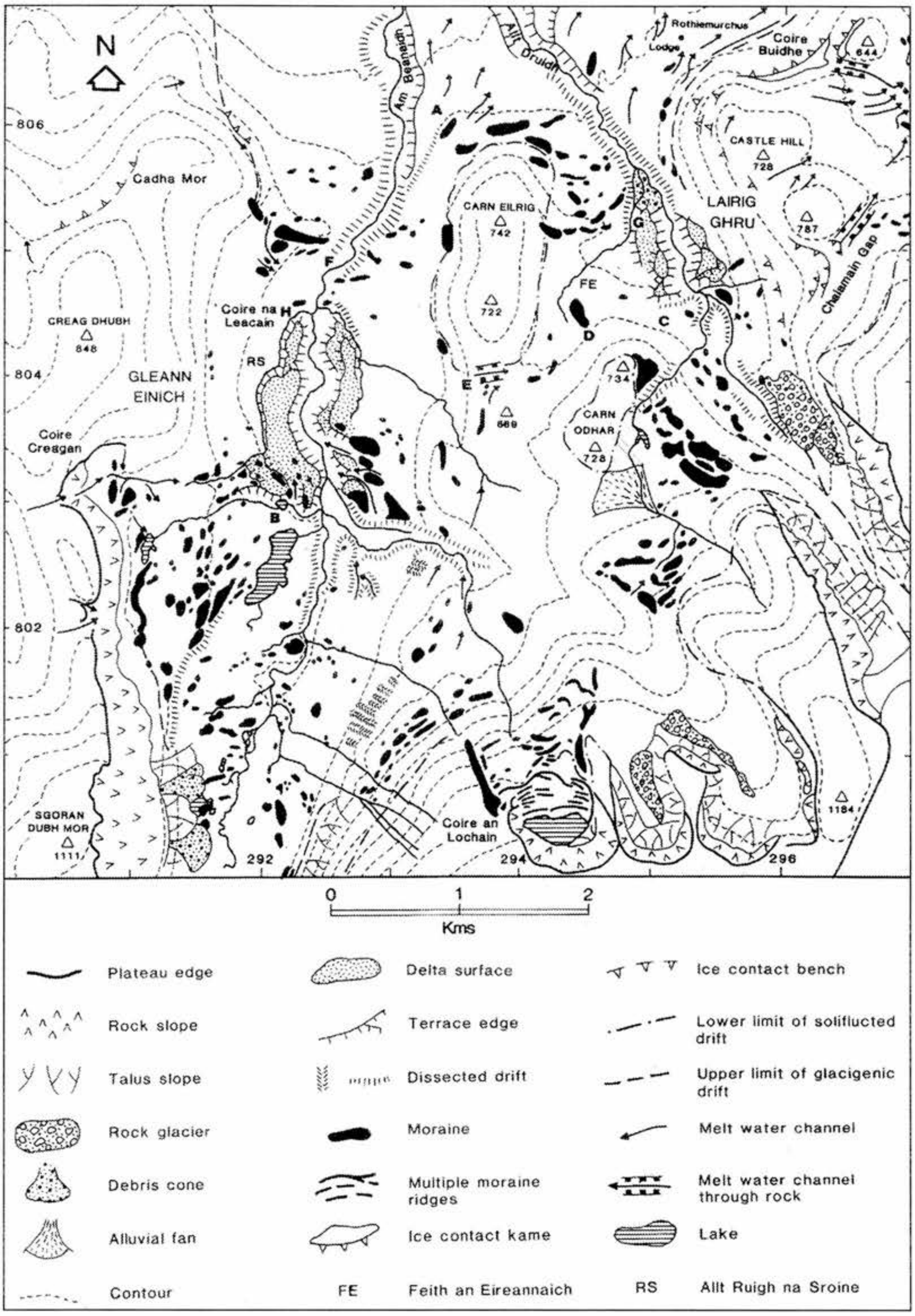


Figure 1.20c: Geomorphology of Glen Einich and the Lairig Ghru as described by Brazier et al. (1998).

3. A readvance of ice in Glen More led to the creation of moraine systems leading into Ryvoan. Ice in the Lairig Ghru separated from the Glen More lobe, creating
4. The final phases of deglaciation in the Cairngorms are difficult to constrain, according to Brazier *et al.* They state that the correlation of moraine limits, such as those proposed by Sissons (1979) is hard to reconcile with the possibility that they may have formed during more than one glacial episode. Final stages of retreat of the Late Devensian can be confused with Loch Lomond Stadial positions, and likewise they point out that many of the high corries contain several terminal moraines, not all of which need to be representative of the Loch Lomond Stadial.

1.6.2e: The Loch Etteridge age constraint

Loch Etteridge (GR: NN/688929) is a large kettle hole, surrounded by a series of kames and eskers (Figure 1.21), and lies some 5 km to the south of the confluence of Glen Truim and the Spey Valley(Walker, 1975). The site is approximately 22 km to the south west of the mouth of Glen Einich in the Cairngorms. At present the loch has provided the only significant dated constraint for the deglaciation of the Spey Valley, and therefore by association it is the only dated constraint for the deglaciation of Glen More. Walker (1975) obtained four radiocarbon dates from the basal section of a core obtained from the deepest part of the infilled kettle at Loch Etteridge, the oldest of these being 13,151 ± 390 ^{14}C BP (15.5 ± 1.1 calendar BP). This date implies that deglaciation of this part of the Spey Valley was complete by this point in the Lateglacial, and by inference the relationship between contemporaneous glacier margins in Glen Einich and Glen More had ceased.

1.6.2f: Summary of Cairngorm research

Over nearly a century of research, there has been broad agreement in the theories of deglaciation in the Cairngorms. Hinxman and Anderson (1915), Charlesworth (1956), Sugden (1968, 1970, 1974, 1980) and Brazier *et al.* (1998) all agree to a great extent over the interaction of Scottish Ice Sheet and Cairngorm ice, and interpretations of the evidence for a still stand, or halt in deglaciation are shared. The one major area of disagreement lies in the interpretation of evidence for the Loch Lomond Stadial in the Cairngorms. Here Sissons, and to an extent authors such as

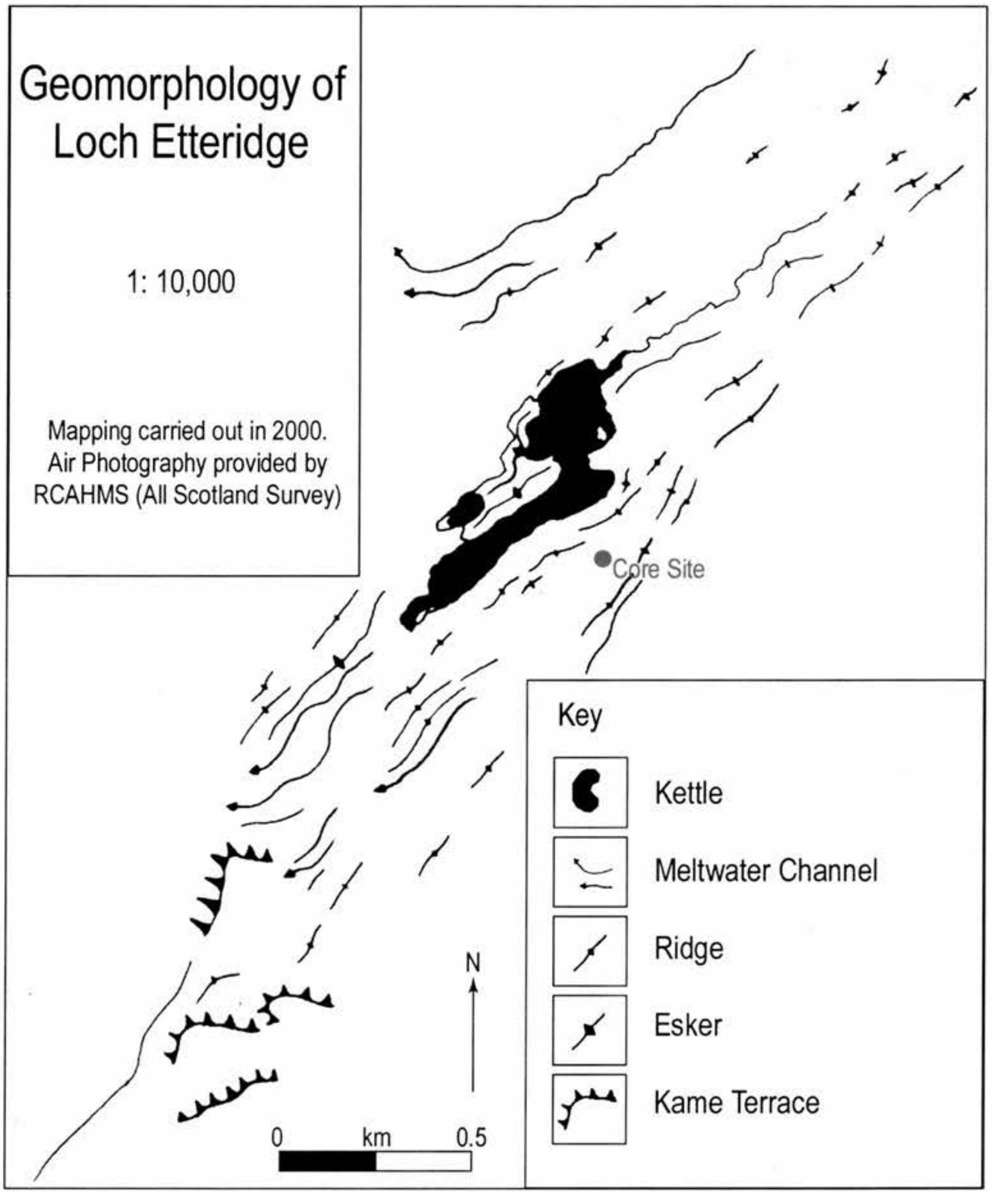


Figure 1.21: General geomorphology of Loch Etteridge, a site marginal to the Spey Valley glacier during its decline in the deglacial. Core site shown (Sissons and Walker, 1974 core site located approx 10 m to the east).

Bennett and Glasser disagree with the consensus view that glaciation was very limited in the massif. Their interpretation of the 'hummocky moraine' landforms as being representative of the Loch Lomond Stadial has influenced the interpretation of the wider glacial story in Scotland, and the rest of Britain. All authors recognise the need for tightly dated constraints in order to better understand the rate and character of deglaciation in the region.

1.7: Aims of this Study

Within the Cairngorms there has been significant, and heated, debate regarding the timing and extent of various glacial phases, in particular the interaction of the major ice sheet systems in the area, i.e. that in the Spey and the Cairngorm ice cap itself. The consensus has emerged however that both the Spey ice from the Scottish Ice Sheet, and the local Cairngorm ice in glens such as Glen Einich and the Lairig Ghru, were contemporaneous for a large portion of their existence. In particular evidence for continuous and contemporaneous stable glacier margin positions in the Lairig Ghru and Glen More, and then in Glen Einich and Glen More during deglaciation, imply strong links between regional, or even hemispheric climate patterns, and the action of regional and local ice masses.

Therefore this work focuses on attempting to provide ages for a single phase of retreat of both the local Cairngorm ice cap, and the Scottish Ice Sheet, represented by the Spey Valley glacier. By mapping limits and dating them using cosmogenic ^{10}Be an absolute constraint for one of the most contentious glacial periods in Scottish deglacial history may be determined. By dating such a limit, chronologies of deglaciation may be constructed for the Cairngorms, and these may be linked to wider Scottish and North Atlantic records of climatic and environmental change, so furthering understanding of the response of these ice sheet systems to fluctuating external variables.

Dated margins representing the deglaciation of the Scottish Ice Sheet in Glen More, and local Cairngorm ice will provide constraints for corrie glaciation in the Cairngorms occurring within, and at higher elevations than the dated margins. The extent of glaciation in the massif during the Loch Lomond Stadial may then be determined.

Chapter 2:

Techniques

Chapter 2: Techniques

2.1: Research Strategy

In order to best complete the aims of this research, a four step strategy was employed:

1. The field area was mapped in detail.
2. A reconstruction was made of specific stages during deglaciation.
3. These stages were dated using cosmogenic surface exposure dating.
4. Constraints for the cosmogenic ages were provided by ^{14}C dating.

2.1.1: Mapping

The field area of the western Cairngorms has been fully and effectively mapped in the past by Sugden (1970) and Sissons (1979), and in part by Young (1974), Bennett and Glasser (1991) and Brazier *et al* (1998). It was felt that remapping of the area was necessary for two major reasons. The first was to produce a representation of the area in line with current thinking in glacial geomorphology and geology. The second reason was to ensure that samples for cosmogenic dating came from features that would provide the most information for the study. In this study 74 vertical aerial photographs at 1:10,000 and 1:24,000 scale were examined with a mirror stereoscope to create detailed geomorphological maps of the western Cairngorm region. Maps created by other workers (Sugden, 1970, Sissons, 1979, Brazier *et al.* 1998, Young, 1974, Bennett and Glasser, 1991) were studied in order to gain an understanding of previous thinking regarding the geomorphological evolution of the Cairngorms. Contentious features or interpretations of landforms were checked in the field.

Moraine, terrace and channel systems were surveyed using tape measure, ranging poles and abney level. Larger scale features such as channel long profiles, lake shoreline and delta altitudes were levelled using a Geodimeter EDM®, and information produced was mapped using Surfer Win32, version 6.04®. Where possible, logs of sedimentary sections were taken in the field. Sections were dug, attempting to cause as little environmental damage as possible, and then cleaned. Photographs were taken in the field, and logs were created, following the standard nomenclature described by the QRA guide to sedimentary interpretation (Jones *et al.* 1999). In some cases, representative sections were sampled in the field for further detailed study in the laboratory.

2.1.2: Reconstruction

Landforms such as moraines were used to delimit former glacier margins. These features were used in conjunction with erosional features such as trimlines, indicating glacier altitudes, and other depositional features such as lake shorelines and deltas. From the correlation of these features, stages of glacier retreat could be reconstructed, whereby glacier snouts in different valleys could be compared, and allocated to particular periods during the decline of the various ice masses.

The geological provenance of erratics in various sections, and on moraine crests gave valuable insights into the source areas of glaciers. The Cairngorm region is almost entirely granite, whereas the Spey Valley is composed of schist material. Sugden (1970) argued that local ice 'flushed' granite from the massif during the later stages of deglaciation, and where schist erratics were present this was indicative of an earlier phase when the area was overridden by the Scottish Ice Sheet from the south west. The locations of former lakes, their longevity and interaction with glaciers was derived from the positions and development of ice contact delta systems in the glens. These were formed by meltwater draining from and around the margins of the glaciers, emptying into lakes occupying the floors of the glens.

2.1.3: Cosmogenic Surface Exposure Dating

Cosmogenic surface exposure dating is a science that has grown in both scope and respectability in recent years. The reasons for this are numerous, though the improved precision of the technique as a result of advances in AMS, laboratory, experimental isotopic physics, and field techniques, has led to its acceptance as a valuable tool for Quaternary reconstruction. Possibly the most prolific area of the science is the application of the technique to the reconstruction of Quaternary ice volumes and extents. Studies have been carried out on a variety of different scales from continental ice sheets (Nishiizumi *et al.* 1991, Gosse *et al.* 1993, McCarroll & Nesje, 1993, Brook at el, 1995a, b, Ivy-Ochs *et al.* 1995, Bierman *et al.* 1999, Zreda *et al.* 1999) to moraine studies (Phillips *et al.* 1990, Gosse *et al.* 1995a, 1995b, Phillips, *et al.* 1996, Ivy-Ochs *et al.* 1999). The precision of such studies is now at such a level that relatively short lived events, such as the Younger Dryas period can be dated effectively (Ivy-Ochs *et al.* 1999). By using multiple samples from individual features such as moraines, precision of data of less than 5% may be achieved (Gosse, *et al.* 1995a).

Because of the ability to produce relatively high precision ages for moraine features, several moraines were sampled with two specific aims in mind. The first aim was to determine if certain moraines were synchronous. This applied to moraines from both the local ice in the Cairngorms, and to the regional Scottish Ice Sheet. Synchrony within the Cairngorms would help to identify stages of deglaciation of the massif, and synchrony of these moraines with the regional ice moraines would enable correlation with broader environmental processes. The second aim was to determine when these stages occurred, and if any correlation could be made with other climatic or oceanic signals present in the Scottish and North Atlantic records.

2.1.4: ^{14}C Dating at Loch Etteridge

As cosmogenic dating is subject to a number of corrections due to production rates, and environmental factors such as horizon and rock surface shielding, and erosion, constraints were needed to provide minimum age limits for the cosmogenic samples. To this end samples were collected for ^{14}C dating from a previously dated site at Loch Etteridge in the lower Spey Valley. Advances in radiocarbon dating required that ages derived from the site in the past (Walker, 1975) must be checked to ascertain their validity.

2.2 Cosmogenic Surface Exposure Dating

Cosmogenic isotopes are produced in a number of minerals by several different methods. This study is concerned with the production of cosmogenic ^{10}Be in quartz. However, the production of other isotopes, such as ^{26}Al, ^{36}Cl, ^{3}He and ^{21}Ne, will be briefly discussed.

2.2.1: Mechanisms of Cosmogenic Isotope Production

The production of cosmogenic isotopes is achieved through two main processes, both involving the incidence of cosmic radiation, firstly at the surface of the upper atmosphere, and secondly at the earths surface. Cosmic rays entering the upper atmosphere create a shower of secondary particles, predominantly neutrons,

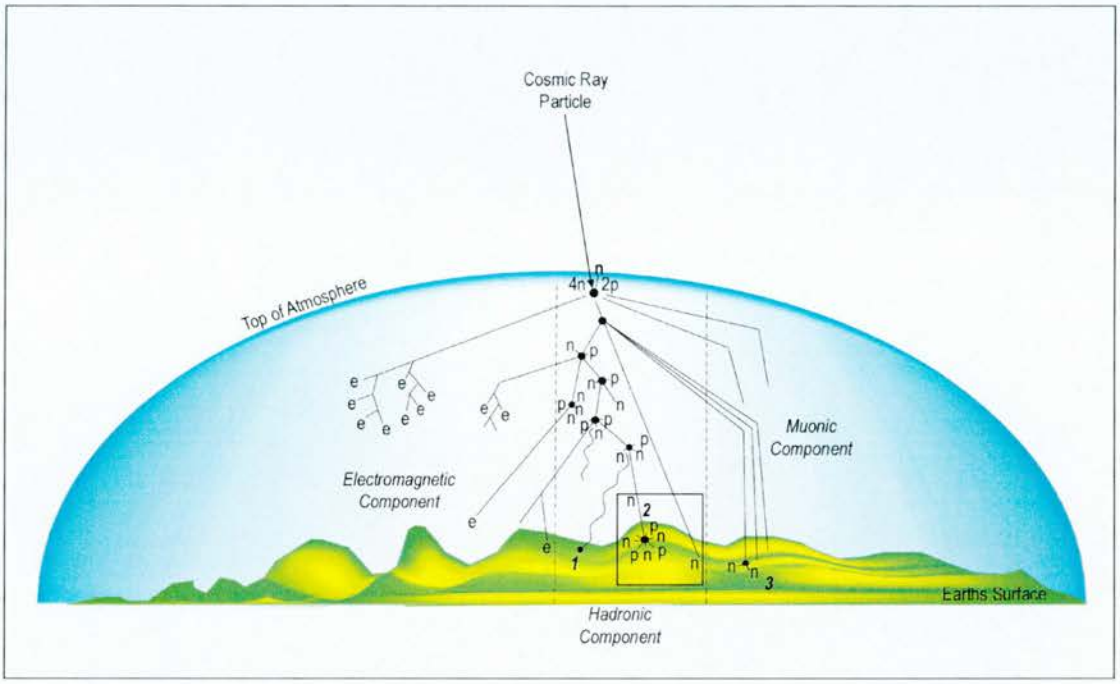


Figure 2.1: The major components of a cosmic ray air shower (cascade), showing secondary particle production (n = neutron, p = proton, e = electron) in the atmosphere and Earth surface (after Gosse and Phillips, 2001). Numbers refer to examples of in situ cosmogenic nuclide interactions: ***1****: ^{35}Cl (n) ^{36}Cl,* ***2****: ^{16}O (n, 4p3n) ^{10}Be,* ***3****: ^{28}Si (n,p2n) ^{26}Al. Boxed area includes reaction for ^{10}Be.*

though muons and protons are smaller components (Figure 2.1). When particles hit minerals within rocks at the earths surface, cosmogenic isotopes are produced by spallation and muon-capture reactions. The isotopes produced by spallation within quartz (^{14}C, ^{10}Be and ^{26}Al), and those in minerals such as olivine (^{3}He ^{21}Ne), are suitable for cosmogenic measurements.

Spallation is the most important of these for ^{10}Be production, both in the atmosphere and in quartz. Spallation within minerals in quartz grains is the primary source of ^{10}Be for surface exposure dating studies. Spallation is a nuclear reaction, in the case of ^{10}Be, ^{14}C and ^{3}He, involving an oxygen atom within a quartz grain. The atom will spall and produce several daughter nucleons within the quartz lattice, one of which may be a ^{10}Be atom. Atmospherically produced ^{10}Be only has a bearing on surface exposure dating studies as a contaminant of samples, through introduction to the earth's surface by precipitation.

Muon capture reactions occur in similar locations to spallation reactions, however the muons produced can be captured by a nucleus, which then decays to form a long-lived or stable isotope. There is evidence to suggest that muon contribution to total production is lower than has been assumed in the past. Early values of 16% of total production (Lal, 1991) have been reduced to around 3% (Stone, 2000).

Once in the quartz lattice at the earth's surface, ^{10}Be is a long-lived isotope, though not totally stable. It has a half life of 1.5 Ma, longer than those of ^{26}Al and ^{36}Cl which have half lives ($T½$) of 0.71 and 0.3 Ma respectively. This makes all three isotopes ideal for the study of many long term geomorphological phenomena. As a result of the process of secular equilibrium (Figure 2.2), the saturation of isotopes within rock surfaces over time due to the balance of build up and decay, means that there are effective temporal limits to the techniques, of 5, 2.5 and 1 Ma for ^{10}Be, ^{26}Al and ^{36}Cl respectively (Ivy-Ochs, 1996).

Production of cosmogenic ^{10}Be can be calculated using equation 1:

$$N = \frac{P}{\lambda + \frac{\rho\varepsilon}{\Lambda}}\left(1 - e^{-(\lambda + \rho\varepsilon/\Lambda)T}\right) + N_0 e^{-\lambda T} \tag{1}$$

where

N is the number of atoms/ gram of SiO_2

N_O is the number of atoms/ gram of the cosmogenic isotope present at the beginning of exposure

P is the local production rate in atoms/ gram/ year

T is the length of time the surface has been exposed

λ is the decay constant of ^{10}Be in 1/yr

ρ is the rock density in g/cm^3

ε is the erosion rate in cm/yr

Λ is the cosmic ray attenuation length in the rock surface in g/cm^2

This equation is greatly simplified when prior concentration of the isotope in the rock surface is zero, and when erosion rate of the rock surface is assumed to be zero.

To account for relative amount of production due to muons, a second scaling factor must be applied using equation 2:

$$M_{\lambda}(P) = M_{\lambda,1013.25} \exp[(1013.25 - P)/242] \tag{2}$$

where

M_λ is production due to muons, scaled by latitude λ in 10° increments (Lal, 1991)

2.2.1a: Atmospheric Thickness and Geomagnetic Shielding Corrections

All surface exposure dating studies rely on the assumption that production rates are constant for the particular latitude and altitude of a sample. Such rates are consequently adjusted from the sea level and high latitude value determined experimentally to the particular spatial location of the sample (Lal, 1991, Dunai, 2000, Stone, 2000). Such adjustment requires a scaling function, and the most commonly employed is that of Lal (1991, Table 2). This function uses a mathematical model which fits average atmospheric pressure as a function of altitude. Pressure therefore varies with altitude and is not constant, though it is now acknowledged that the model is over-simplistic as a representation of actual global conditions. Local annual average atmospheric pressure variations are common, particularly over continental interiors (where pressure is often higher than the norm for such latitudes), and in oceanic areas such as Iceland (where low pressure dominates). It is not known how significant these anomalies are when averaged over timescales of 10^3 to 10^5 years.

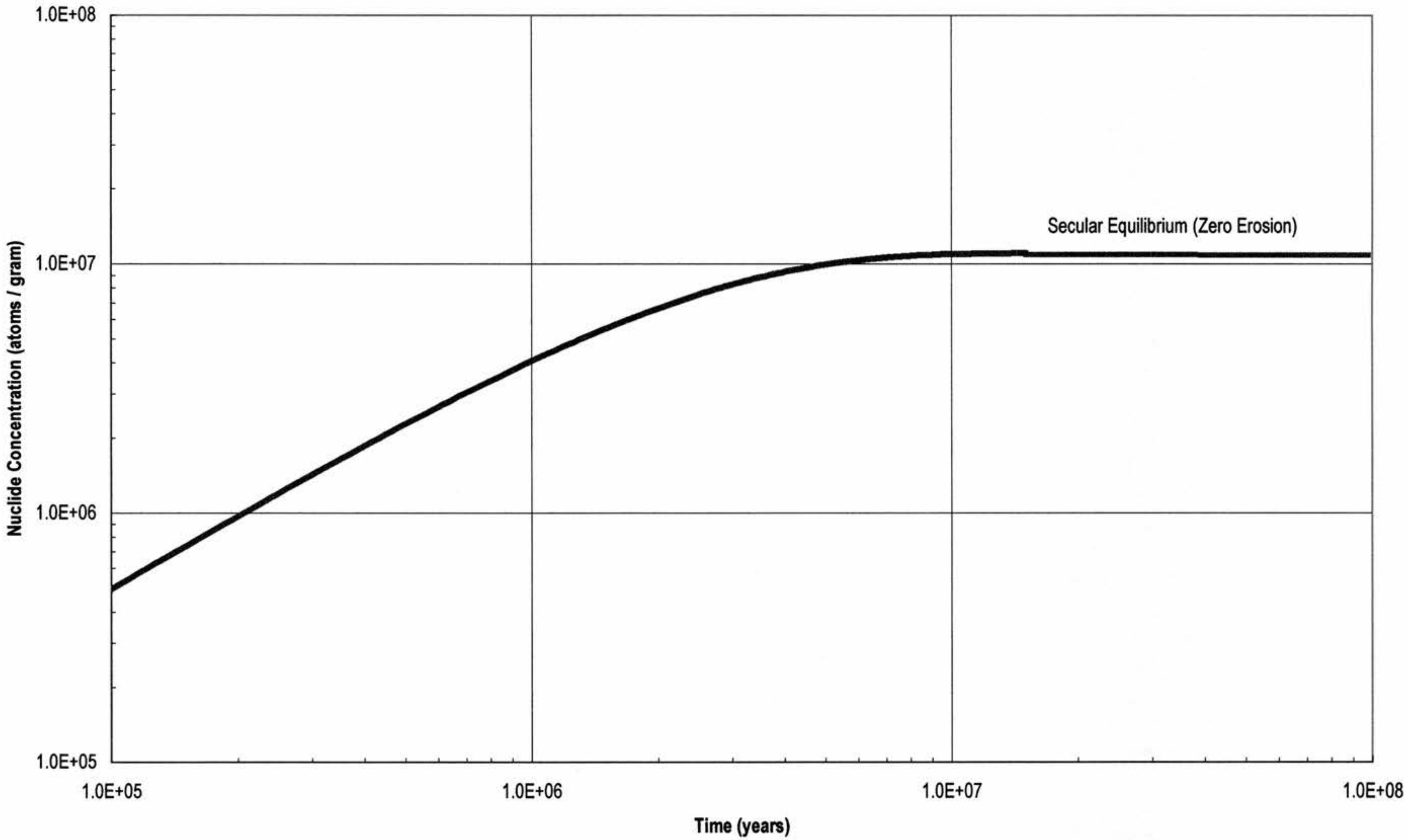


Figure 2.2: The increase in 10Be concentration (atoms / g) as a function of time (years). Secular equilibrium is reached at around 1.0 x 107 years for a zero erosion model of nuclide accumulation.

Isotope production increases with altitude, as the atmosphere acts as a partial shield to incoming cosmic rays though as a corollary to this, atmospheric pressure variations through time will act as a proxy altitudinal variation. Therefore it is argued that one must determine the fluctuations of local pressure systems through time, if one is to make accurate exposure age estimates based on a constant production rate for a certain point in space (Stone, 2000). Using averaged annual pressure data for the last 50 years, Stone (2000) argues that areas such as Scotland experience lower pressure on average for latitudes 54 - 58° N, resulting in an increase in production rate (at sea level) of +2%. Such estimations do not yet take into account pressure variations over longer timescales, nor do they include changes of altitude as a result of uplift, subsidence or isostasy.

2.2.1b: Rock Thickness Corrections

Production of cosmogenic isotopes in minerals varies downward through any rock surface, as attenuation of cosmogenic rays occurs with depth. Lal (1991) argues that production rate varies exponentially with depth from the surface. However Masarik and Reedy (1995) assume no shielding in the rock surface to an equivalent depth of 12 g/cm^2 (or 4.6 cm depth) in granite, (assuming the density of granite is 2.6 g/cm^3, and the attenuation length is 157 g/cm^2). The Masarik and Reedy method therefore assumes a lower sensitivity of the production rate to shielding at the rock/ air boundary as the correction is only applied to the thickness of sample over 4.6 cm. The correction also reduces sensitivity of the sample to coverage by snow or soil. As such it is a more sensible depth correction method to employ where such factors are unknown. More traditional 'straight exponential' thickness corrections produce a purely mathematical reduction in production with depth. Where snow or soil coverage (or even shielding from surrounding vegetation) may be a factor, this is excluded, and forms no part of the calculation, though if it were to be included, then significant alterations in production rate with depth would be made. The Masarik and Reedy correction therefore overcomes this problem by assuming negligible shielding in the upper surfaces of the rock sample, and consequently if a layer of snow were to cover the sample, this would be factored into the low sensitivity rock / air interface, producing only small variation in production rate.

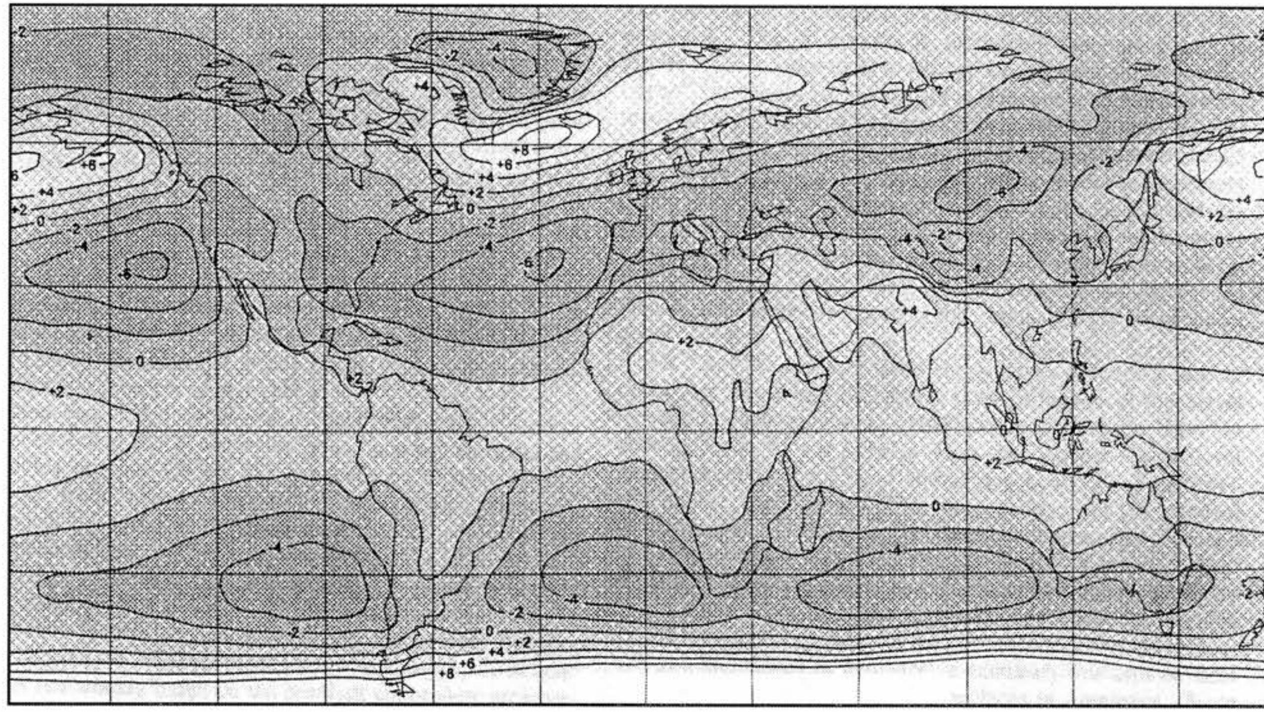


Figure 2.3: Corrections to production rates proposed by Stone (2000). Corrections are expressed as percentage deviations from the scaled production rates proposed by Lal (1991), which are based on a uniform sea level atmospheric pressure of 1013.25 hPa. Stones corrections are based on mean pressure calculated over 50 years, and are positive above areas of low mean pressure, and negative over areas of high mean pressure. As a secondary consideration they are based on geographic, rather than geomagnetic latitude. In this way Stone proposes that long term fluctuations in geomagnetic dipole position are averaged. Corrections are not shown on this figure for the Antarctic region, where Stone argues that corrections are of the order of 25 to 30% (Stone, 2000).

Depth dependent production can be calculated as follows:

$$P_z = P_o \exp(-z / z^*)$$

(3)

where

P_z is the production rate at depth z (atoms/gram SiO_2/year)

P_O is the production rate at the surface (atoms/gram SiO_2/year)

z^* is the distance from the surface whereby production is decreased by $^1/_3$ (i.e. where $z^* = \Lambda/\rho$)

Λ is the neutron attenuation path length

ρ is rock density

For this study granite was assumed to have a density of 2.6 g/cm³ and an attenuation length of 157 g/cm² (Masarik and Reedy, 1995).

2.2.2: Environmental Factors

2.2.2a: Horizon Shielding

Any sample at the Earth surface receives cosmogenic rays from any direction that is exposed to open sky. Therefore a sample on a totally flat plateau, open to the full hemisphere of the heavens, would have zero environmental shielding. This is usually not the case when studying samples in deglaciated terrain. Therefore a correction must be made depending on how much of the open sky is obscured by surrounding topography, and also by the surface topography of the sampled rock surface itself. The shielding for each individual sample is calculated by measuring angles of elevation of the horizon or nearest obstacle for every 15° of the azimuth.

Horizon shielding was calculated from:

$$SC = 1 - \frac{1}{360°} \sum_{i=1}^{n} \Delta\phi_i \sin^{m+1} \theta_i$$

(4)

where

θ is the angle to the horizon

ϕ is the azimuthal angle

n is the number of obstructions

m is an experimentally determined constant (taken as 2.3 ± 1.2)

2.2.2b: Rock Surface Shielding

Shielding of the surface of a sample is difficult to quantify with certainty. It is often impossible to know whether or not a sample has been covered by soil, vegetation (such as moss) or snow for any length of time during its exposure since deposition. Corrections can be made, based on the rock thickness corrections already described, assuming depths of coverage by various media. Snow coverage can be assumed by introducing a correction based on an extra thickness added to the rock surface, averaged over time. Briner *et al.* (2002) employ a 4% correction derived from modern snow and rime ice coverage of boulders in the Ahklun Mountains in Alaska. This can only be an estimate as palaeo-snow coverage depths are unknown. Likewise soil and vegetation coverage effects can be estimated using similar corrections. A further complication exists when attempting to assume coverage by surrounding vegetation such as trees. More often than not the vegetation history of a sample site is incomplete or non-existent, though assumptions can be made based upon local palynological records, or peat/ soil stratigraphy.

2.3: Sampling

These corrections may provide some estimation of total shielding, but there is no way of knowing whether they are accurate (Figure 2.4). How long does snow lie at the site, and how long has it lain during winter seasons over preceding millennia? How dense was the forest cover? There can be no realistic and satisfactory answer to these questions, but such corrections usually affect production rates in samples by one or two percent.

For the purpose of this study no snow or vegetational corrections were employed, as sampling was restricted to large, upstanding boulders of greater than 1.5 m as this reduces the likelihood of long term coverage of their upper surface by earth, till or long-lying snow. Boulders close to, or on ridge crests were chosen to minimise the risk of post depositional movement, either through settling, or rolling off the ridge itself as a result of mass movement processes such as solifluction. Once the boulder had been chosen, the upper surface was sketched, showing the main orientation of faces and surface angles. This surface was labelled for ease of identification during the next stage of sample preparation. Where possible the top surface of the sample was removed intact using a mallet and sharp chisel. The sample was then photographed *in situ* both close up, and from a distance allowing some impression of landscape context to be seen.

Figure 2.4: Sampling concerns. Environmental shielding from surrounding slopes is measured at 15° intervals to the height of the horizon, the amount of sky obscured is calculated using equation 3 Other possible sources of shielding include vegetation, snow and soil cover. Each of these is difficult to quantify without detailed knowledge of the post-depositional environment, however they are acknowledged as having an effect on any age estimates. Erosion due to spalling of rock surfaces, granular disintegration or dissolution, is often evident in the soil surrounding the boulder. Theoretical estimation of effects of erosion on exposure ages can be made using equation 5.

This study is concerned with the deglaciation of the Cairngorms, and therefore morainic ridges were the primary locations for the collection of rock samples from erratic boulders. The moraines selected for sampling were chosen for a number of reasons.

1. The landforms are distinctive and are likely to represent glacier margins.
2. The moraines identify stable glacier positions.
3. The moraines formed by different local and regional glaciers are contemporaneous.
4. One of the moraines (Glen Geusachan) represents a controversial limit that has been the subject of extensive debate concerning the glacier age and extent during the Loch Lomond Stadial (Sissons, 1979, Sugden, 1970, 1974 Bennett and Glasser, 1991).
5. All of the moraines appear to record the final large-scale still stand of the local ice in the Cairngorms, and regional ice in the Spey Valley.

2.4: Assumptions Made When Using ^{10}Be as an Age Estimation Tool (Figure 2.4)

2.4.1: Production Rate

Cosmogenic production rate calculation is fundamental to age estimation. Published production rate values of ^{10}Be in quartz vary considerably, from 4.74 atoms g^{-1} yr^{-1} (Clark *et al.* 1995) to 6.4 atoms g^{-1} yr^{-1} (Brown *et al.* 1991). Such variation in the calculation of production rates is significant. The result is that depending on the choice of production rate, final age estimations of any given sample may vary by thousands of years (Table 2.1).

P (at/g/yr)	Researcher	Location
6.03 ± 0.29	Nishiizumi et al (1989)	Sierra Nevada
6.4 ± 1.5	Brown et al (1991)	Antarctica
6.13	Nishiizumi et al (1991)	Antarctica
6.0 ± 0.3	Nishiizumi et al (1996)	Water target
5.75 ± 0.24	Kubik et al (1998)	Koeffels
5.1 ± 0.3	Stone (2000) from Brown (1995)	
5.97	Masarik & Reedy (1995)	Theoretical
5.1 ± 0.3	Stone (2000)	Skye
5.02	Barrows et al (2002)	Corrected from standardisation error

Table 2.1: Production rates employed by various researchers.

Scaling factors are applied to production rates to compensate for differences in latitude and altitude (or latterly atmospheric thickness). A problem arises because the use of different production rates requires different methods of scaling, depending on how the original rate was calculated. Older production rates (Nishiizumi, 1991, Brown *et al.* 1991, Clark *et al.* 1995) employ the scaling factors published by Lal (1991), which use a production rate scaled to high latitude and sea level, and include production of ^{10}Be in quartz as a result of spallation and a high muonic component of 15.6%. More recent production rates (Stone, 1998, 2000, Kubik *et al.* 1998) require scaling factors with production dependent upon both spallation and a low percentage of total production as a result of muon capture of around 0-5%. By employing a 3% muon contribution at sea level, then all published production rates converge on 5.1 ± 0.3 atoms g^{-1} yr^{-1} (Stone, 1999, Gosse & Phillips, 2001). Theoretically therefore all production rates, when scaled to sea level and high latitude using this method, can be compared, and discrepancies in published rates can be resolved (Gosse and Phillips, 2001).

As research progresses it is likely that future estimates of production will require recalculation of data already published. This will not mean that current and previous work is redundant, but interpretations based upon outmoded production rates may well have to be reassessed. For the purpose of any study it is clear that the production rate must be unambiguous, and the assumptions for the latitude and altitude of the site over the period be made explicit.

2.4.2: Prior Exposure

In the simplest situation it is assumed that any given sample has no prior exposure history, or if it has been previously exposed, then subsequent erosion has been sufficient to remove all cosmogenic isotopes before re-exposure. The source of erratic boulders may present difficulties here. It is assumed that erratics are primarily subglacial in origin, moving to the glacier margin englacially or subglacially, their surfaces being eroded as they travel. Therefore it is best to sample striated or faceted boulders, where the subglacial mode of transport to the glacier snout is clear. However, a significant proportion of morainic material is produced by paraglacial processes, such as rockfalls onto the glacier surface. Such material may then be incorporated into the glacier, and transported englacially, or it may be carried on the glacier surface to the margin. Exposure during transport is not a significant issue, as transport times at the most are of the matter of centuries, but exposure on the valley side, prior to the rockfall taking place, may well produce erroneous ages if such boulders are sampled. Porter (*pers. comm.*) suggests that the chances of such a prior-exposed boulder surface being sampled if the source is known to be from rockfall, are 1:6 (the study was based on the assumption that boulders were cubic in shape). Again the use of multiple samples and careful sample selection are employed to reduce the chance of sampling a boulder with inherited cosmogenic isotopes.

Studies have been carried out using samples collected from exposed bedrock surfaces. Such investigations are extremely useful when attempting to determine exposure histories or erosion rates of surfaces. However they are less applicable to exposure age estimation. The nature of striated bedrock surfaces ensure that there is often insufficient surface removal (by erosion) to zero the exposure record. For studies attempting to determine an exposure age the use of erratic boulders is more commonplace (Gosse *et al.* 1995a, 1995b, Ivy-Ochs *et al.* 1996, Ivy-Ochs *et al.* 1999).

Landscapes modified by warm-based glaciers are among the most suitable environments for the use of ^{10}Be dating, as they satisfy one of the main assumptions made, that of zero prior exposure.

2.4.3: Additional Variation in ^{10}Be Levels Due to Other Factors

Following on from this, most studies assume that there has been no addition of ^{10}Be to the surface since exposure. This may occur in the form of precipitation, containing meteoric ^{10}Be. Likewise no loss of ^{10}Be is assumed to have occurred by leaching or diffusion of the isotope from quartz grains.

2.4.4: Sample Movement

The movement of boulders since deposition can affect the amount of cosmogenic isotope build-up in their upper surfaces. Just as with problems associated with lack of knowledge related to surface shielding, it is difficult to determine if a boulder has moved since deposition. Such activity may change the orientation of a boulder, such that what was once the upper surface is now the lower, or *vice versa*. If such movement has occurred, a falsely young age for deposition will result. Unless it is obvious in the field if this has happened (and if so, the boulder is rejected as a potential target), the only way of discovering whether post depositional movement has occurred is through the interpretation of AMS results, which demonstrate that the boulder does not exhibit a consistent age. Such problems are reduced by sampling from multiple boulders on each landform. This produces its own problems when potentially erroneous ages appear, as choices need to be made to determine which of the samples is anomalous.

2.4.5: Erosion

Surface erosion is the final, and possibly most important factor that may affect cosmogenic isotope values, and thus the inferred age of a sample. There are empirical methods for estimating erosion rates, based on the comparison of concentrations of two or more cosmogenic isotopes within a sample. If isotopes with different half lives are used, it is possible to compare the value of each isotope, using the decay rate as a constraint to the possible absolute maximum concentration, and thus define the difference between the expected, and actual concentrations. This method inevitably precludes the use of the maximum concentration of isotopes as an exposure age study, as one creates the circular argument that erosion is constrained by age, and age by erosion rate.

By far the most common method used, is simply to include an estimated erosion rate (McCabe, *et al.* 2001 etc). In this way an envelope of possible maximum and minimum exposure ages may be determined. One can never be sure if this envelope is correct. However studies of the sampled rock surface itself, combined with published work can help mitigate against erroneous erosion rates. A relatively simple study that can be undertaken, particularly in areas where such features are common, is to measure the elevation of quartz veins above the surrounding rock ground mass. As veins are far more resistant to weathering than polymineralogic material, as is especially the case in granites, they are likely to survive exposure intact for much longer periods. Assuming the rock surface was lowered to the same level by glacial erosion, veins will stand proud of more easily weathered ground mass. Various studies have also been undertaken worldwide to estimate surface denudation rates, and those that were deemed relevant to this study, in terms of lithological, climatic and environmental similarity have been consulted (Mottram, 2001).

Erosion rate estimates produce an envelope of ages as follows:

$$T = \frac{-1}{\lambda + (\varepsilon / z^*)} . \ln\left(1 - \frac{N^*.\lambda + (\varepsilon / z^*)}{P}\right) \tag{5}$$

where

T is exposure age dependent upon erosion

P is production rate corrected for sample thickness and environmental shielding

N is atoms/g/yr corrected for sample thickness and environmental shielding calculated at site*

λ is the decay constant of the radionuclide

ε is the rate of erosion in cm/kyr

z is attenuation length (157 g / cm^2) divided by rock density (2.6 g / cm^3)*

The envelope of possible age estimates can therefore be bounded by employing a zero erosion rate, to give minimum possible age, and then using a high rate of erosion to estimate the maximum age. By working through the equation using higher and higher rates of erosion, a point is reached whereby infinite ages are reached, as production reaches its secular equilibrium. At this point the model is predicting the *creation* of more rock than can possibly have been removed by erosion, and therefore gives an effective upper limit to erosion.

2.5: Laboratory Protocols

Throughout the whole of the sample preparation procedure several protocols were followed to ensure contamination was kept to a minimum (Figure 2.5). All containers, pipettes and instruments were rinsed in dilute HNO_3, and then triple rinsed in 18MΩ H_2O. All Teflon beakers were washed for 2 hours in dilute HCl, dilute HNO3 and finally 18MΩ H_2O on a hotplate at 80°, and stored dry and capped. Sample preparation was carried out in a 'clean-air' positive pressure laboratory, and at all times workers wore gloves, lab coats and 'clean overshoes'.

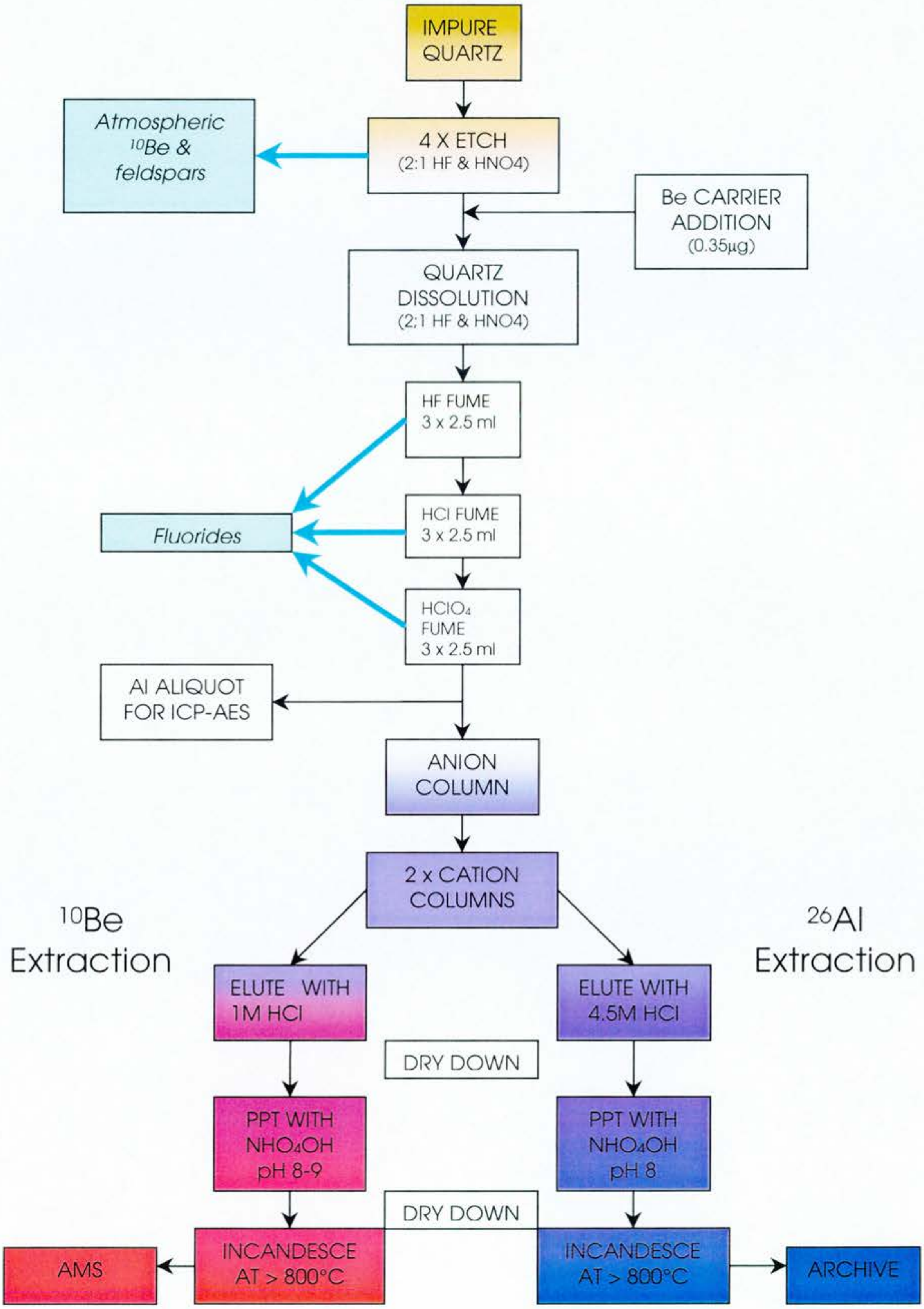


Figure 2.5: Generalised flowchart of laboratory protocol highlighting main stages in the production of clean samples of ^{10}Be and ^{26}Al.

2.5.1: Crushing and Separation

Each rock sample was cleaned of lichen, then dressed using a circular saw to produce a sample of as near uniform thickness as possible. This thickness was measured, both at maximum, and average for the whole sample. The rock was then crushed using a jaw crusher to 250 μm. After sieving and repeated crushing to reduce as much of the initial rock to the required grain size, the sand was washed thoroughly with 18ΩM H_2O to remove fines, and then dried. In some cases the sample was placed in a dilute (1M) HCl solution to digest any remaining organics over 24 hours. Once the sample had been dried, density separation was carried out using LST. The aim of this process was to separate as much of the quartz as possible from the other minerals present in the granite sand (predominantly micas and feldspars). Once again the sample was thoroughly washed and dried.

2.5.2: Quartz Etching

For the accelerator to obtain a sufficiently accurate detectable measurement, a minimum number of atoms of cosmogenic beryllium and aluminium are required. This value can be estimated by knowledge of the exposure time (estimated / projected), and the latitude and altitude of the sample site.

By calculating the expected production rates, and then using an exposure age estimate the mass of pure quartz required can be ascertained. Due to the lack of naturally occurring ^{10}Be in Quartz, a known volume of standard ^{9}Be carrier is used to calculate the ratio between this and the unknown concentration of ^{10}Be. The minimum ratio of the number of stable ^{9}Be atoms to the number of radiogenic Beryllium must be around 10^{-12} to 10^{-13}. Any lower than this will possibly be below the detection levels of the accelerator.

Etching of the separated quartz removes any atmospheric ^{10}Be, produced by spallation in the upper atmosphere and washed to earth by precipitation. This step in processing is vital, as any additional ^{10}Be in the sample will lead to erroneous results. To achieve this, the sample was etched using concentrated Hydrofluoric (HF) and Nitric (HNO_3) acids (2:1) in solution in water. This step was repeated four times, the sample was dried, and checked under a microscope for any mineral grains resistant to the etch. These were either removed by hand picking, or the etch was repeated until the sample contained only pure quartz.

2.5.3: ^{9}Be Carrier Addition and Dissolution

Once a sufficient mass of pure quartz separate has been produced by etching, a precise measurement (E x 10^{-5} grams) of the mass of each sample was made both prior to and after addition of a known volume of ^{9}Be Carrier. Carrier solutions contain precisely measured concentrations of ^{9}Be atoms, and are therefore used for calibration at the AMS where the ratio of ^{9}Be to ^{10}Be can be compared. The sample and carrier (both in a Teflon beaker) were transferred to a hotplate at 80° C, and covered in a 2:1 solution of concentrated HF and HNO_3, until the sample was completely dissolved. This process may take days or up to a month, depending upon the size of the sample.

2.5.4: Fuming

Once dissolved the sample was dried to form a precipitate, and then fumed three times with 2.5 ml HNO_3 to remove any SiF_4. The process was repeated with HCl and finally with $HClO_4$. This sequence removed any remaining fluorides not destroyed by dissolution. Especial care was taken at this stage as $AlCl_3$ may be lost through volatilisation if the hotplate temperature exceeded 183° C. Such a loss results in errors in the total Al measurement at the later ICP-AES stage.

2.5.5: Al Aliquot

The Al aliquot, though not strictly necessary for preparation of ^{10}Be samples, is still extremely useful as a guide to the dissolved mineral chemistry of the cleaned precipitates. The sample was taken up in a small volume of HNO_3 and transferred to a 100 ml conical flask, which was then filled with 18MΩ H_2O. A further 10 ml sub sample was removed and transferred to a second 100 ml flask, which was also filled. The second flask therefore contained a known sample solution of 10%. A sub sample of this second flask was sent for ICP-AES analysis.

When carrying out age estimations using ^{26}Al the ICP results indicated whether an Al carrier needs to be added, if there is insufficient ^{26}Al in the original sample. However as this project is only concerned with results from ^{10}Be measurements, ICP results were used only as a guide for concentrations of interfering isotopes, along with confirmation of the appropriate carrier volume. Following extraction of the ICP sample, the rest of the sample was transferred to a clean Teflon beaker, and reduced on a hotplate to around 2 ml.

2.5.6: Anion Column

The anion column stage is not strictly necessary for most samples, unless they happen to contain a large concentration of iron minerals such as magnetites as inclusions within the quartz grains. The results of such inclusions are seen as a strong yellow colouration once the sample has been taken up in dilute HCl following the drying down after the aliquot stage. If not removed, iron can overload the later cation column stage, rendering it unable to separate Al and Be atoms. Column conditioning was undertaken before the first sample run and between each sample (using 20ml columns i.e. 1cv=20ml. Columns used were Biorad analytical grade Ag50W X 8 cation exchange resin, H^+ form). Three column volumes (cvs) of 18MΩ H_2O were eluted, removing any Zn or Cl complexes present, followed by 3 cvs 9N HCl to remove any remnants of the previous sample. The sample was placed on the column in solution (<5ml 9NHCl), and the first 5ml eluted from the column was collected with the conditioning volumes, labelled and archived as the 'sample volume'. 2 cvs of 9N HCl were eluted through column and collected in a savillex beaker. This volume should contain Be and Al.

To elute any remaining elements such as iron minerals, 4 cvs of 0.5N HCl were placed on the column. These were collected, bottled and labelled 'Bottle 1'(cvs 3-6).

CVs	Acid Used	Volume	Solution Contains:
Column conditioning	18MΩ H_2O	60 ml	Zn, Cl complexes
Column conditioning	9M HCl	60 ml	Previous Sample
Sample volume	9M HCl	5 ml	Nothing
CVs 1-2	9M HCl	40 ml	Al and Be
CVs 3-6	0.5M HCl	80 ml	other anions

2.5.7: Sample Clean Up

The sample was fully dried down on a hot-plate at 150^{0}C. 2-3ml of HCl was then added to the sample to ensure it was all in solution. Using ammonia solution (NH_4O_4) and pH paper the solution was brought to a pH of 8, to allow precipitation of Al and Be. This also allowed any left over Boron (^{10}B), Mg and Ca to go into solution, if left to stand for 24-48 hours (Ivy-Ochs 1996). The sample was then centrifuged (3000 rpm, 15 min), and the supernatant decanted and archived.

2.5.8: Optional cleaning procedures.

The sample was then assessed, primarily on the basis of its colour. If any yellow was seen (indicating a high Fe content) the sample was rinsed with 18MΩ H_2O, and an HCl fume was carried out. Following the fume HCl (<1N) was added, prior to cation column stage.

2.5.9: Cation Column

The function of the cation exchange columns is to separate the $Al(OH)_3$ from the $Be(OH)_2$ using different acid strengths to elute the Be, and then the Al. The 20 ml columns were conditioned with 3 cvs of 18MΩ H2O, 3cvs 9M HCl, 3cvs 4.5M HCl and finally 3cvs 1M HCl. This column-conditioning sample was archived, as remnants of the previous sample may be recovered from it if later mistakes occur. The sample was placed on the column in the <1N HCl solution added in the last stage. Once collected, the Al and Be samples were dried down to c. 2ml, and the two elutant volumes labelled and archived. Now the Al and Be are separate, great care was taken to avoid cross contamination due to the air movement of the fume hood.

If the sample was >40 g, the two fractions were dried down. The Al fraction was put aside and the cation column procedure was repeated for the Be fraction only (including column conditioning). This removed any excess Al potentially contaminating the Be fraction. The 80 ml Al elutant was added to the original Al fraction in the Teflon beaker from the first column run.

CVs	Description / Strength	Vol	Solution Contains:
Column conditioning	3 cvs of H_2O	60 ml	**complexes**
Column conditioning	3 cvs of 9M HCl	60 ml	**Previous sample**
Column conditioning	3 cvs of 4.5M HCl	60 ml	**nothing (equilibrate)**
Column conditioning	3 cvs of 1M HCl	60 ml	**nothing (equilibrate)**
Sample Volume	<1M HCl	<5 ml	**nothing**
cvs 1-3	Bottle 2 (1M HCl)	60ml	**elutant**
cvs 4-11	Savillex beaker(1M HCl)	160ml	**Be**
cvs 12-15	Bottle 3 (1M HCl)	80ml	**elutant**
cvs 16-19	Savillex beaker (4.5M HCl)	80ml	**Al**

2.5.10: Precipitation of Separated Be and Al

Once the sample has completely run through the columns, the Teflon beakers containing the separate $Al(OH)_3$ and $Be(OH)_2$ aqueous solutions were transferred to a hotplate and dried down at 90°. The separate samples were taken up in 2 ml of 1M HCl and transferred to clean disposable centrifuge tubes. NH_4OH was added to each until the $Be(OH)_2$ and $Al(OH)_3$ were precipitated (at pH 9-10 and 8 respectively) and left for 24 hours for any remaining Boron to go into solution. The tubes were centrifuged for 10 minutes at 3000 rpm, and the supernatant decanted and archived. The precipitate was rinsed and the supernatant decanted and archived.

2.5.11: Incandescence

The $Be(OH)_2$ and $Al(OH)_3$ precipitates were placed on a heater block and dried at 70° C. Once the hydroxides had formed pellets these were transferred to quartz vials, and the heated over a flame until they were incandescent. The vials were held in the flame until it was certain the whole sample had incandesced. The vials were allowed to cool, and then were capped, placed in labelled packaging, and sent to the AMS to be pressed into targets.

2.6: Summary of Rates and Corrections Used in the Cairngorms

2.6.1: Production Rate

For the purpose of this study the rate proposed by John Stone (2000) of 5.1 ± 0.3 atoms g^{-1} yr^{-1} has been employed. This rate was calculated by rescaling all published calibration data allowing for reduced muon capture production. The rate is therefore somewhat lower than earlier published rates such as those used for previously glaciated terrain in the Sierra Nevada, of 6.0 atoms g^{-1} yr^{-1} (Nishiizumi *et al.* 1996). The site chosen for Stones' study lay on the northwestern coast of Scotland, at the same latitude and similar altitude to the Cairngorm samples presented here. Likewise the published age range of Stone's samples covers the expected range of those in the Cairngorms, reducing the impact of other complicating factors, such as geomagnetic field variations, when calculating the exposure ages.

2.6.2: Latitude/ Altitude Correction

Scaling factors for latitude and altitude were taken from Stone (2000), who employs corrections in terms of atmospheric pressure variation rather than altitude, and also allows for relatively lower contribution to production in terms of muons. Recent work (Gosse & Stone, 2001, Gosse & Phillips, 2001) has suggested that corrections used for isotope production by muon capture may be too high (they suggest a muonic component of ~ 2% of total ^{10}Be production at sea level, rather than 15.6%, which is the figure used for existing altitude corrections).

If average global sea level pressure is taken as 1013.25 hPa, for the altitude of the Cairngorm samples, local pressure (determined by equation 6) varied between 964.95 and 946.61 hPa. All samples were scaled to a latitude of 57° N, and functions for relative production of muons and spalled atoms were applied to each sample.

$$P(z) = P_s \exp\left\{-\frac{gM}{R\xi}\left[\ln T_s - \ln\left(T_s - \xi z\right)\right]\right\} \tag{6}$$

where:

P_S = Sea level pressure (1013.25 hPa)

T_S = Sea Level Temperature (288.15 K)

ξ = Adiabatic Lapse Rate (dT/dz = 0.0065 K m^{-1})

M = constant molar weight of air

g = Acceleration due to gravity

R = Gas Constant

gM/R = 0.03417 K m^{-1}

2.6.3: Sample Thickness Correction

The Masarik and Reedy (1995) thickness correction was performed on the samples. This correction assumes no shielding in the rock surface to an equivalent depth of 12 g/cm^2 (or 4.6 cm depth) in granite (density of granite 2.6 g/cm^3, and the attenuation length is 157 g/cm^2).

2.6.4: Distant Shielding Correction

Measurements of environmental shielding were taken in the field for each sample at 15° intervals, and then the shielding correction was applied using equation 4.

Corrections varied from 3% on the Glen Einich lateral moraine samples to 0.3% in Glen Geusachan.

2.6.5: Blank Correction

Measured atoms/g^{-1} data for each sample were corrected by subtracting the appropriate measured blank atoms/g^{-1} to allow for any input of ^{10}Be to the sample during processing.

2.6.6: ICP-AES Blank Measurement

As pipette measurement of the absolute quantity of carrier added to each sample is imprecise, the ICP-AES data were used as the correct totals for calculation. This gave accurate amounts of carrier added to each sample to a precision of $10e^{-8}$ mg.

2.6.7: Error Propagation

Error in all measurements was carried through the calculations, when attempting to determine the accuracy of the age estimate. For the purpose of calculation, error in the age is derived from uncertainty in the measurement of the AMS. Uncertainty in other site-specific factors is accounted for by applying the various corrections already detailed. Counting accuracy on the AMS varied between 4.1 and 8.7%, with a mean AMS error of 6.41%. Added to this is a 5% reproducibility error (calculated as a percentage of the total AMS error). The error correction is applied to the site specific production rate for each sample, rather than the atoms/g^{-1}.

2.7: Loch Etteridge Radiocarbon Constraint

With recent advances in AMS ^{14}C, and the possibility that the original date from Loch Etteridge may have been affected by mineral carbon contamination (Walker, *pers comm*), it was decided that a second study should be undertaken to check this vital constraint to deglaciation by the Scottish Ice Sheet in the region.

The infilled Loch Etteridge sedimentary basin was mapped out using a Dutch gouge, to determine the deepest point from which a core could be removed. Stratigraphy of the section was logged using techniques described by Jones, *et al.* (1999). Once the deepest point was located, the core was removed using a 10 cm diameter Russian corer. As no pollen analysis was to be carried out on the core, only the 50 cm basal section was removed. The base of the section was clearly delineated by the presence of fluvial gravels, similar to those described by Walker (1975).

In the laboratory samples were taken from this section for Loss on Ignition testing, to determine the relative amount of organic material down-core. Sampling for loss on ignition requires the removal of material at 1cm intervals downcore, each subsample then weighed to a precision of e x 10^{-4}. Subsamples are then dried for a period of up to 48 hours, and then reweighed. Finally samples are placed in an oven and fired at 550°C for 4 hours and then left to cool from 925°C in order to burn off any organics. The subsamples are weighed for a final time. This final 'ignition weight' is plotted against dry weight to produce a % loss of organic material on ignition signal. Samples were also taken at 1 cm intervals for tephrostratigraphical analysis. Each subsample being wet sieved through a series of gauze mesh sieves of 75 μm and 24 μm. Fractions recovered from this sieving are then examined under a microscope for presence of tephra shards (Figure 4.4).

The sample to be sent for AMS radiocarbon dating was removed from the lowest section of the core containing organic sediments. This sample was prepared employing established techniques (Smart and Frances, 1991), and sent to the NERC Radiocarbon Laboratory at East Kilbride.

Chapter 3:

Glacial Geomorphology of the Western Cairngorms

Chapter 3: Glacial Geomorphology of the Western Cairngorms

(See Geomorphology Map Insert to rear)

3.1: Introduction

The morphology of the Cairngorms is a result of the continuous and discontinuous operation of geomorphic processes since the exposure of the surface of the granite batholith some 65 Myr ago (Glasser and Bennett, 1996) (Figure 3.1). The overall morphology of the area can be described as a rolling plateau with incised valleys (Sugden, 1968, 1970, Rea, 1998). In the glens there is considerable geomorphic complexity related to the activity of wasting glaciers.

This chapter describes the geomorphology of the western Cairngorms, in particular three primary areas of the Glen More basin, northern flanks of the Cairngorms and Ryvoan; Glen Einich and the Lairig Ghru; and finally Glen Geusachan, (Map insert and Figure 3.2). Following on from this is a reconstruction of the pattern of deglaciation, fitting the various landforms into a relative chronological framework. It is this relative chronology that is necessary in the cosmogenic isotope analysis.

3.2: Glen More, the Northern Flanks and Ryvoan

3.2.1: Glen More

The Glen More basin (Figure 3.3) lies to the north of the Cairngorm massif, its north eastern limit defined by the hills of Craiggowrie (686 m), Creagan Gorm (732 m), Meall a' Bhuachaille (810 m) and Creag nan Gall (622 m). To the northwest lies the Spey Valley and the hills of the Monadhliath. The south western limit of the basin is marked by an abrupt transition from the basin floor to the hills surrounding the mouth of Glen Einich; Creag Dubh (848 m) and Carn Eilrig (742 m). From the air the basin appears as an eastward extension of the Spey Valley, a 'bite-shaped' lowland, bounded on all sides by hillslopes. The area encompassed by the basin is a triangle 7 km along its southern base, and 6 km along its north western and north eastern sides.

Immediately to the west of the southwestern boundary of Glen More basin are three hills, Ord Ban (428 m), Kennapole Hill (382 m) and Torr Alvie (358 m). Each of these appears smoothed on its south western slope, and steeper on the north eastern side. All display smoothed bedrock at their summits, and reflect overriding ice from the south west down the Spey Valley (Hinxman and Anderson, 1915, Young, 1974).

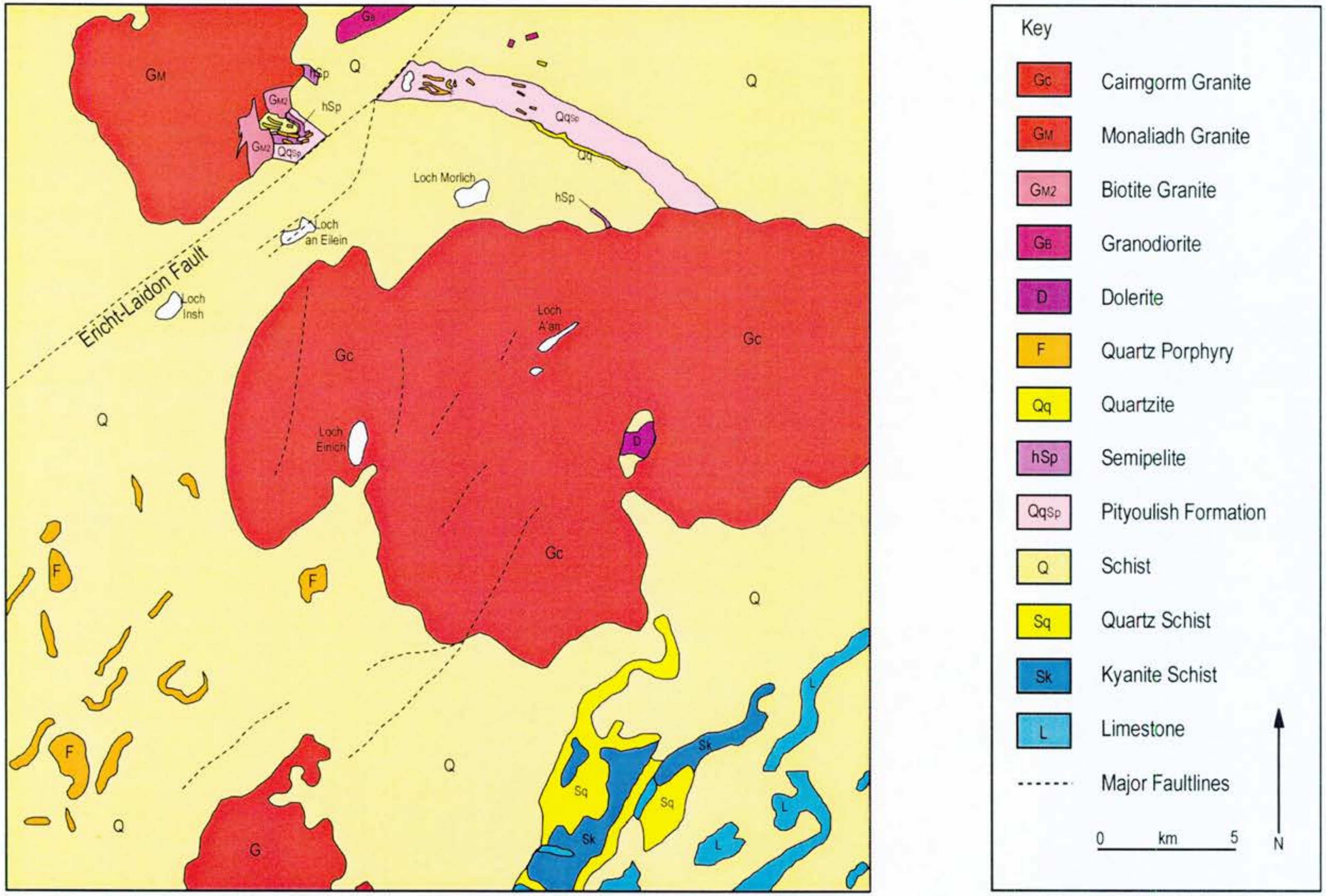


Figure 3.1: Geology of the Cairngorm region (after Geological Survey of Great Britain (Scotland) Sheet 64 (!964) and Sheet 74E (1993).

Figure 3.2: Satellite image of the western Cairngorms and the southern part of Glen More Basin. Contours are in 50 m intervals.

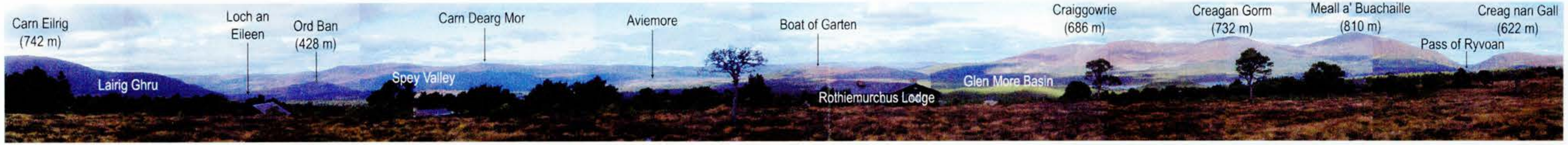


Figure 3.3: Panoramic view of the Glen More Basin, seen from Rothiemurchus Lodge (GR NH952068)

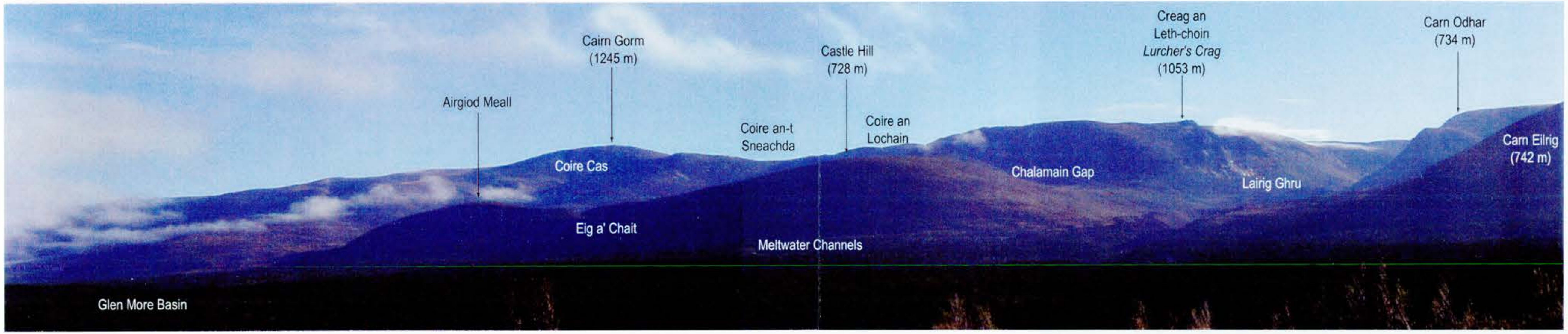


Figure 3.4: Panoramic view of the Northern Flanks of the Cairngorms, seen from Whitewell (GR NH917091)

The floor of the basin itself is undulating, and rises gently from the Spey valley in the west to the foot of the Cairngorm slopes in the east and south. Despite extensive commercial forestation, the land surface morphology can be seen clearly both from the ground, and from above. Extensive networks of channels and ridges are evident across the basin, coupled with sinuous ridge systems west of Loch Morlich (Young, 1974). Towards its eastern limit the floor of the basin is marked by a series of channels, draining from south to north along the northern flanks of Carn Eilrig, Castle Hill, Airgiod Meall, and into Ryvoan Pass. To the west of these channels, the ground surface morphology becomes less ordered, and ridges are interspersed with frequent infilled lake depressions.

3.2.2: The Northern Flanks

3.2.2a: Lower meltwater channels

The area from the west at Cadha Mór to the Pass of Ryvoan in the east, is characterised by a particular suite of landforms. The bedrock morphology is that of slopes of 10 to 20° rising to the south, however upon this has been superimposed a network of channels and ridges along the entire length of the northern flanks of the massif (Figure 3.4 and 3.6). Typically these channels are flat bottomed, and between 10 and 50 m wide, separated by ridges of up to 20 m in height. In cross section they appear as a series of steps, separated by the ridges. Consequently the upslope side of each channel is higher than the downslope side. They dip to the east at angles of 2 to 3°, and many are continuous for several kilometres (Figure 3.5). In plan form they are predominantly linear, however they exhibit arcuate meanders along their lengths. They do not support significant present day fluvial drainage, and are clearly relict features. Exposed sections are uncommon, but where the interior structure of the ridges is visible (particularly in road cuttings), sequences are generally clast supported, and composed of coarse gravel and sand deposits. Sections are dominated by sub rounded cobbles (up to 20 cm in diameter), with sands and gravels forming the majority of the matrix. Little silt or clay material is present. The floors of the channels themselves are covered by thick peat deposits, and many support bog lakes.

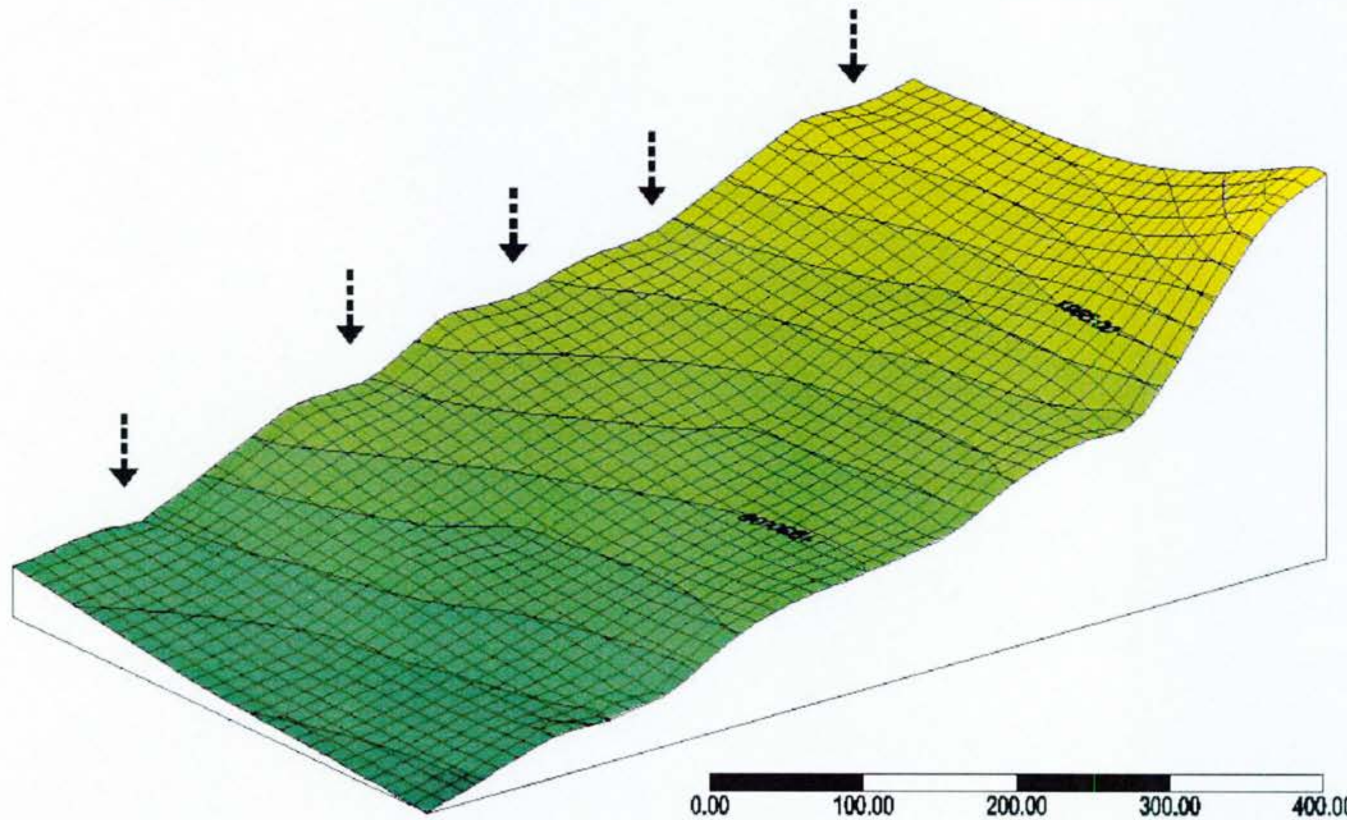


Figure 3.5: (DEM created using survey data from Geodimeter EDM, and SurferWin32 Version 6.04. Interpolation of points completed using Kriging method). DEM of channels on the northern flank of Airgiod Meall. Five major and two minor channels can be identified on the eastern (left) side of the figure. The interpolation used in the software tends to smooth the surface topography, however the overall cross sectional profile, and eastward-dipping trend of all channels can be clearly seen.

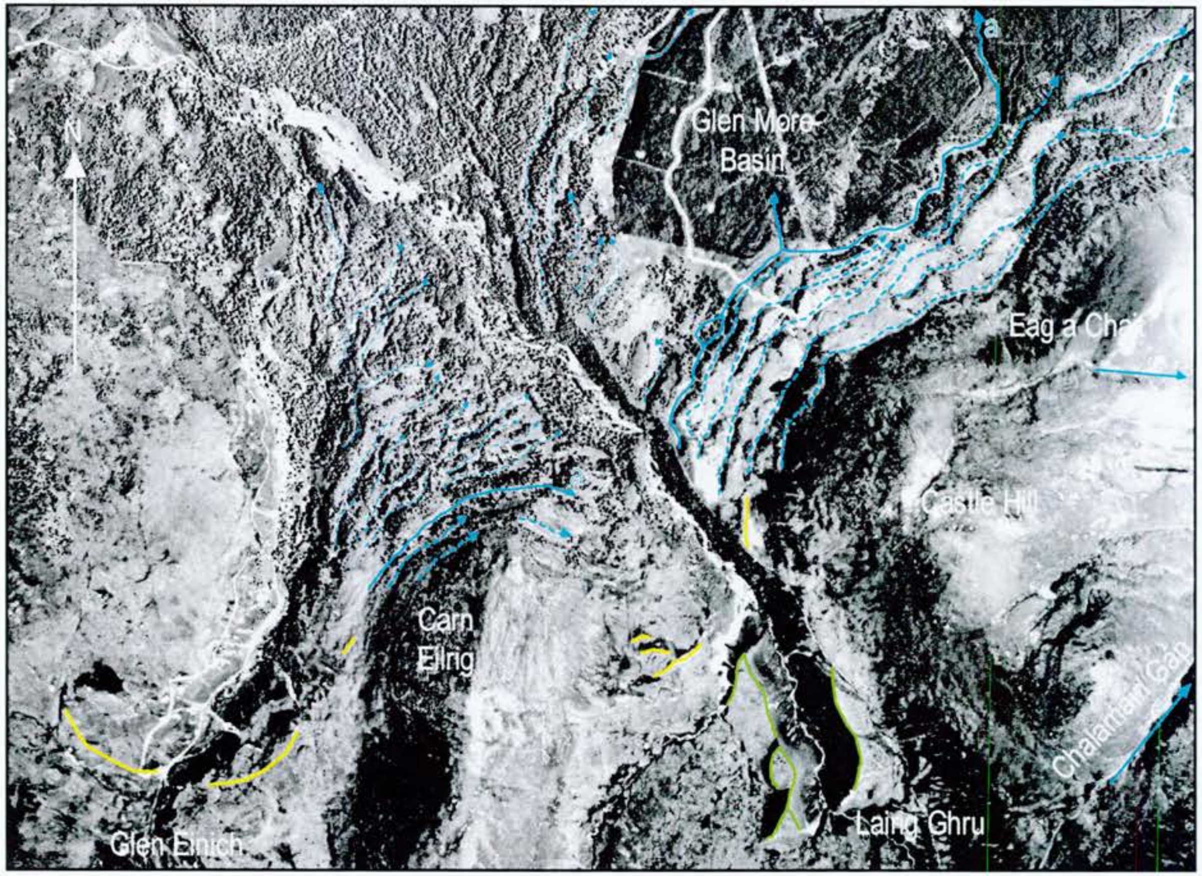


Figure 3.6: Cairngorm northern flanks and Glen More Basin. Moraine sequences (yellow) in Glen Einich are seen to be younger than those in the Lairig Ghru by their relationship to meltwater channels. Channel 'a' is that most closely linked to the Glen Einich moraines by field relations, and is seen to be 45 - 50 m lower than the Lairig moraines. This channel is also seen as the last large scale feature of this kind- those below it are significantly less well-developed (fine dashed blue arrows). The 'Glen Einich Stand' of the Glen More Lobe represents the last period of snout stability of the Scottish Ice Sheet in the Spey Valley. What follows is a period of rapid retreat, and final collapse of the Spey Glacier. Delta sequences in the Lairig Ghru (green) are therefore representative of an earlier ice-dammed lake system than that in Glen Einich.

At their eastern extent the channels become broader (up to 100 m wide), and they change direction from flowing predominantly northeast, to a more northerly, downslope direction. Beneath Creag nan Gall this sequence of channels flows into the Ryvoan Pass, and finally north eastward towards Abernethy.

3.2.2b: Upper meltwater channels

Higher above the floor of the Glen More basin lies a second series of channels with different characteristics. The upper series comprises the incised rock channels of the Chalamain Gap, Eag a' Chait, Caochan Dubh a' Chadha, Eag a' Garbh Coire, and Carn Lochan na Beinne. Between Creag a' Chalamain and Creag an Leth-choin (Lurcher's Crag) at an altitude of 700 m, the Chalamain Gap cuts through the north western spur of Lurcher's Crag (Figure 3.7). This impressive feature is some 100 m wide at the crests of the cliffs on either side, but less than half this width on the floor of the channel. The cliffs rise 20 – 30 m on either side, and the channel itself is around 400 m long, dipping to the north east at an angle of 3-4° though rising c. 15 m from the western intake, before continuing downhill (Young, 1974). The Eag a' Chait channel is a similar though somewhat smaller channel, running east south eastward between Airgiod Meall to the north, and Castle Hill to the south west. The altitude of the intake is 568 m, and the channel experiences a fall of around 10 m over its' 350 m length. At the outflow end, the channel continues eastward for 1500 m, in a less confined, though still deep and steep sided valley. The course of the modern Caochan Dubh a' Chadha stream follows this valley, until it turns abruptly northward at its confluence with a similar channel draining from Coire Cas. To the south east of the summit of Carn Lochan na Beinne another channel flows northward from a kame terrace above Lochan na Beinne. This channel falls 100 m to an extremely broad (100 m) kame, south east of Creag nan Gall. This kame leads into another channel which dissects the col between Creag nan Gall and Màm Suim, falling 160 m towards the Abernethy Forest. From here the channel enters the complex channel systems draining Strath Nethy (Young, 1974).

Figure 3.7: The Chalamain Gap looking north eastwards from GR. 963050. Postglacial slope mass movements have obscured much of the southern (right) side of the channel with talus. However the northern side is still clearly a deeply incised face. The channel has a maximum depth of 30 m, and though it slopes to the north, the intake rises from the Lairig Ghru, before plunging downward through the channel towards the Caochan Dubh a' Chadha, indicating formation subglacially as ice-directed channels (Sugden, 1970, Young, 1974).

3.2.3: Glen Einich

Glen Einich extends northward from the Monadh Mór area of the western Cairngorms. The glen is 3 km wide at its widest point, and is some 8-9 km long from the trough head in the south to the Glen More moraine in the north. The southern part of the glen is a flat bottomed, U-shaped trough bounded by steep, 400m cliffs to the west, and a series of corries to the east draining the western flank of Braeriach (1296m). The head of the glen at Coire Odhar displays a steep backwall which rises 460 m above Loch Einich to the low, flat summit of Carn Bàn Mór (1052 m). The mouth of Glen Einich is bounded on either side by the summits of Crèag Dubh (848m) to the west and Carn Eilrig (742m) to the east.

3.2.3a: The Glen More moraine

A large ridge lies at the northern end of Glen Einich at its confluence with Glen More (Figure 3.8). It crosses the mouth of Glen Einich from GR920053 to GR931049. It is concave in form and extends as a broad arcuate ridge, curving into the 1 km-wide valley mouth. It stands some 70m above the valley floor at its crest. It has a steep distal slope of 30 - 36° in Glen Einich, and a shallow proximal slope of around 10 - 20°, extending northwards around 1 km into Glen More. Ridges and channels occur on the upper surface of the moraine, along with large boulders on its crest overlooking Glen Einich. In the west, a channel cuts the ridge, and slopes into the Glen at an angle of 3°, where it leads into terrace fragments. The ridge is composed of matrix supported, unsorted diamict, containing clasts of both the local Cairngorm granite, and schist clasts, typical of Spey valley deposits (Golledge, 2002). The matrix coarsens upward, with the base primarily formed of boulders in a clay and sand matrix, and the upper sections are comprised of sand and gravel. Throughout the section are large subangular clasts, up to 1 m in diameter. The proximal slope of the ridge surface is dissected by the major east-west flowing channels that are common elsewhere along the northern flanks of the Cairngorms (Young, 1974). These features are typically 10 - 20 m wide, flat bottomed, and bounded by ridges, typically with asymmetric cross sections, standing some 5 - 15m above the floor of the channels. The channels exhibit a longitudinal gradient of around 2 - 3°, dipping to the east, and are continuous along the northern flanks of Carn Eilrig.

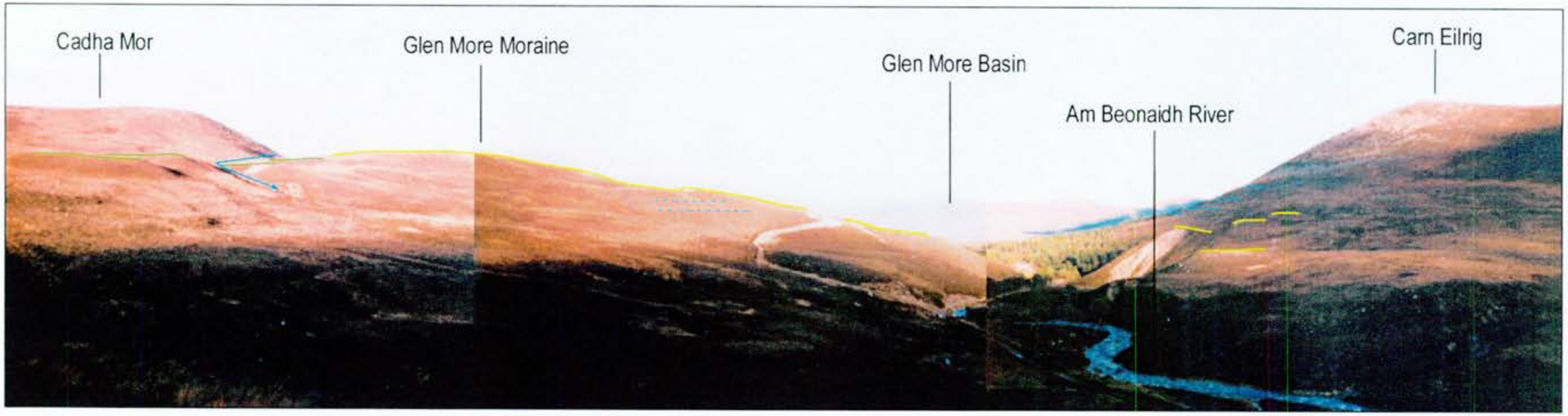


Figure 3.8: Glen Einich looking north toward the Glen More moraine, from the top of the valley floor lacustrine delta set. The crest of the moraine, where discernible (yellow) dips into the centre of the glen, where it has been incised by the Am Beonaidh. On the eastern side of the glen the moraine is discontinuous. Delta sets (green) associated with a meltwater channel (blue arrow) draining into the palaeo lake from Glen More, can be seen in the west of the photo. Other meltwater channels, draining around the flanks of the Glen More lobe, can be seen on the lower slopes of Carn Eilrig. Two low stands of the palaeo lake are shown by a pair of shorelines (dashed blue lines) on the distal slope of the Glen More moraine. These probably represent the lowest stand of the palaeo lake.

3.2.3b: The Glen Einich marginal moraines

The middle of Glen Einich is crossed by a ridge, breached by the Am Beanaidh river. The ridge rises some 40 m above the level of the present river course, and extends discontinuously across Glen Einich for 2 km, from the edge of Coire Creagach to the lower slopes of Carn a' Phris-ghiubhais (Figure 3.9). Distal slopes maintain angles of 25 - 35°, and are bounded to the north by terraces. To the south of the ridge are a number of partially infilled lakes, both immediately behind the ridge, and also to the east. The lower distal parts of the ridge are typified by stratified coarse and medium gravel lenses interspersed within diamict, which can be seen where the Am Beanaidh has incised the ridges at GR: 925030. The ridge has a somewhat concave profile up valley.

On the western side of Glen Einich are a series of stepped ridges, each sloping northwards at angles of 2 - 4°. Partially infilled lake hollows occur between some of the ridges. All of these features support large granite boulders on their crests (Figure 3.11). The outer and westernmost of these features forms a long sinuous ridge, roughly 6 m in height, and with distal slopes of 20 - 30° and proximal slopes of 10 - 20°, which extends for around 500 m from the southern edge of Coire na Saobhaidh to the east of Lochan Beanaidh. Downslope of the uppermost ridge are further ridges which are less well defined, deposited at roughly 30 m intervals. At their most northerly extent, a series of flat topped terraces occur immediately outside the outermost ridge, aligned to channels sub – parallel to the ridges and running downslope from Coire Creagach.

3.2.3c: The Lake Terraces

Between the Glen Einich and Glen More ridges are two terraces (Figure 3.11, 3.12). The western terrace forms a flat topped feature, 1.5 km long by 400 m wide, standing up to 45 m above the glen floor, and its upper surface dipping at roughly 1.5° to the north. The top of the terrace is around 5 m lower than the level of the Glen Einich moraine ridge along its southern edge, and lower also than a series of terrace remnants in the mouth of Coire Creagach. The eastern terrace is less well defined, but maintains a reasonably flat upper surface. It is 1 km long from north to south, and though less distinct than the western feature, it is some 300 m wide. This

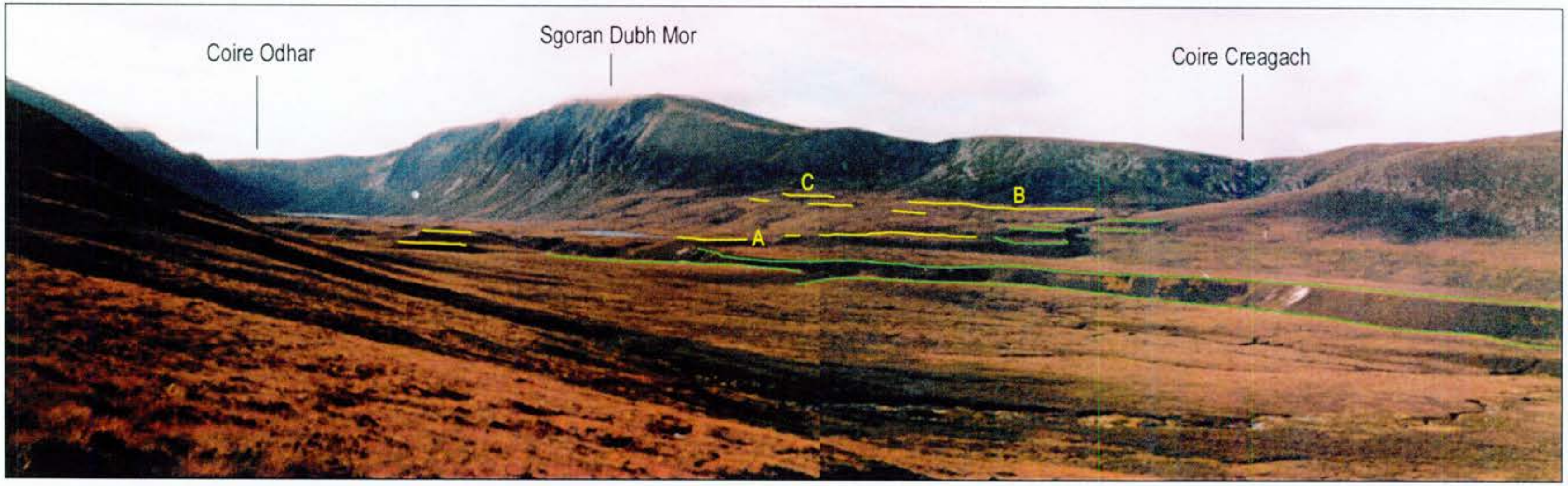


Figure 3.9: Glen Einich, looking southwest from beneath the southern flank of Carn Eilrig. Lacustrine delta sets (green) can be clearly seen on the valley floor, and downslope from Coire Creagach. The Glen Einich terminal moraine (A) lies across the floor of the glen to the south of the valley floor deltas. The lateral and retreat moraine complexes (B & C) can be seen to the south of the delta sets draining Coire Creagach.

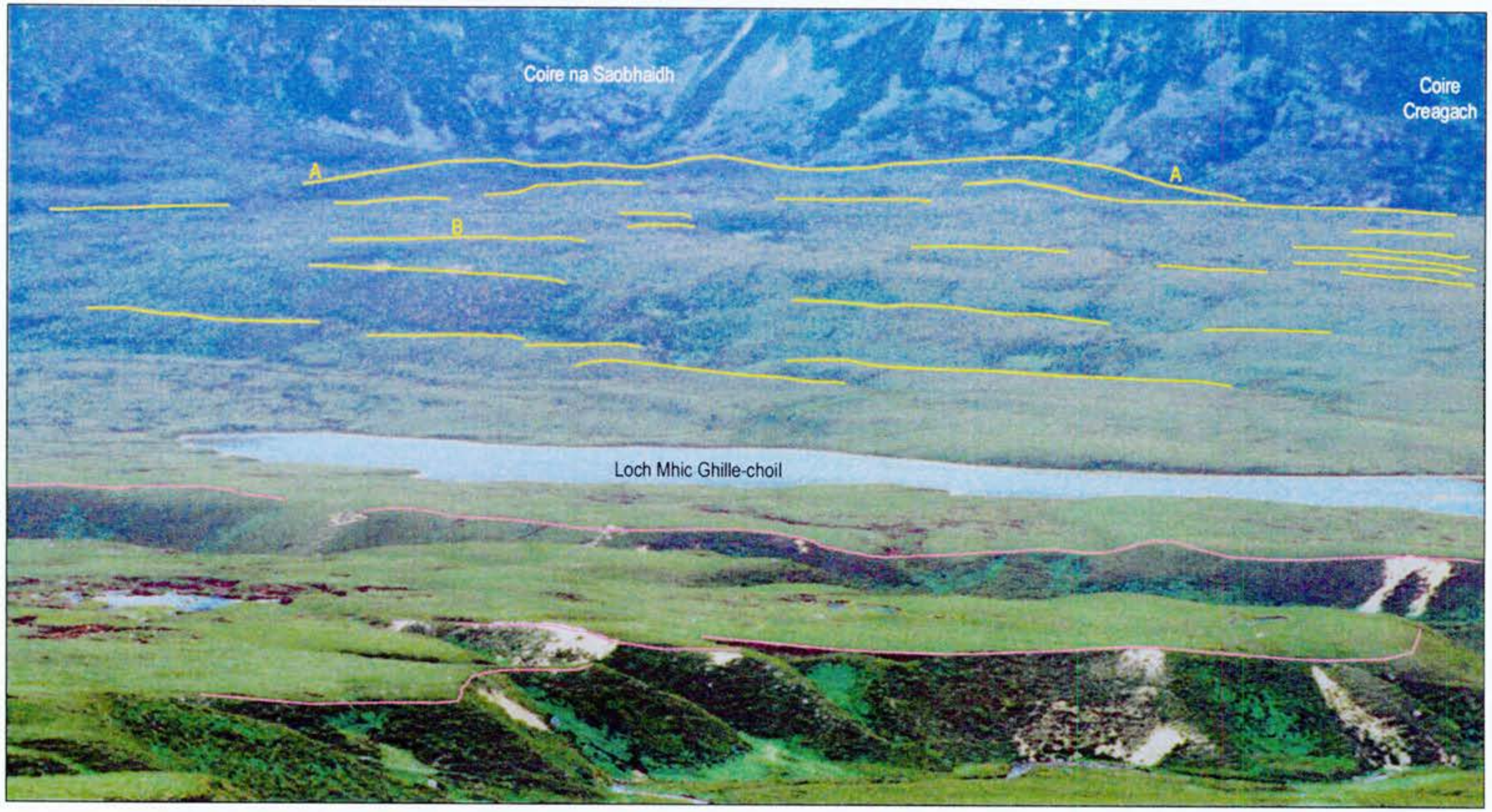


Figure 3.10: Glen Einich Glacier lateral moraines (yellow). The outermost moraine (A) supported sampled boulders CE6 and CE9, and is the most prominent of the moraines in the sequence, being 2-3 m high, and displaying a clear asymmetric cross profile along its entire length. The recessional moraines within limit A are smaller in size, being on average 1-2 m high. Moraine B supported sampled boulders CE7 and CE8. The eastern delta sets can be clearly seen in the foreground (pink). (Photo Nick Golledge, copyright British Geological Survey, 2002)

eastern terrace is deeply dissected by the Am Beanaidh. In the north both terraces are composed of finely laminated silts, clays and sands, topped with coarse gravel topsets (Figure 3.13, 3.14 and 3.15). The figures at both coarse (Figure 3.13) and fine scales (3.14) illustrate the complex nature of the stratigraphy of these delta sets. The composition of both terraces coarsens to the south. A survey of the sediments revealed a rhythmically banded series of silt/clay couplets (Figure 3.13 and 3.14), overlying diamict (Brazier *et al,* 1998). Evidence for lacustrine deposition can be found in dewatering structures and microfaulting within the clay/silt couplets. The rhythmites display numerous dropstones, and also evidence of microtectonism (Golledge, 2002).

A related piece of geomorphological evidence is to be found on the col between Glen Einich and the Lairig Ghru. A channel has been deeply incised into the southernmost portion of the flank of Carn Eilrig, draining from Glen Einich, into the Lairig Ghru. This channel is associated with lacustrine deposits on the western side of the Lairig Ghru (Figure 3.15d).

3.2.3d: The lake shorelines

Two parallel features run north-south for over 2 km along the flank of Carn Eilrig at 510 m and 520 m elevation, highlighted by a clear transition from subaerially formed periglacial features such as solifluction lobes and frost shattered faces, to relatively uniform hillslopes. The break in slope is horizontal. On the western side of Glen Einich, a second pair of discontinuous horizontal features can be seen close to the western terrace at 480 m and 500 m (Figure 3.16). Though less obvious than the features beneath Carn Eilrig, these features, along with those on the eastern side of the valley have been levelled using a Geodometer EDM, and all are horizontal along their entire lengths. At the northern end of these western features, a series of terraces descend from the channel incised into the Glen More moraine. The upper terrace surface is at the same level as the lower shoreline. Further terraces can be seen outside the eastern edge of the Glen Einich moraine. Their upper surfaces correspond with those in the north west. There are also two lower shoreline features evident on the southern flank of the Glen More moraine at 459 m and 466 m.

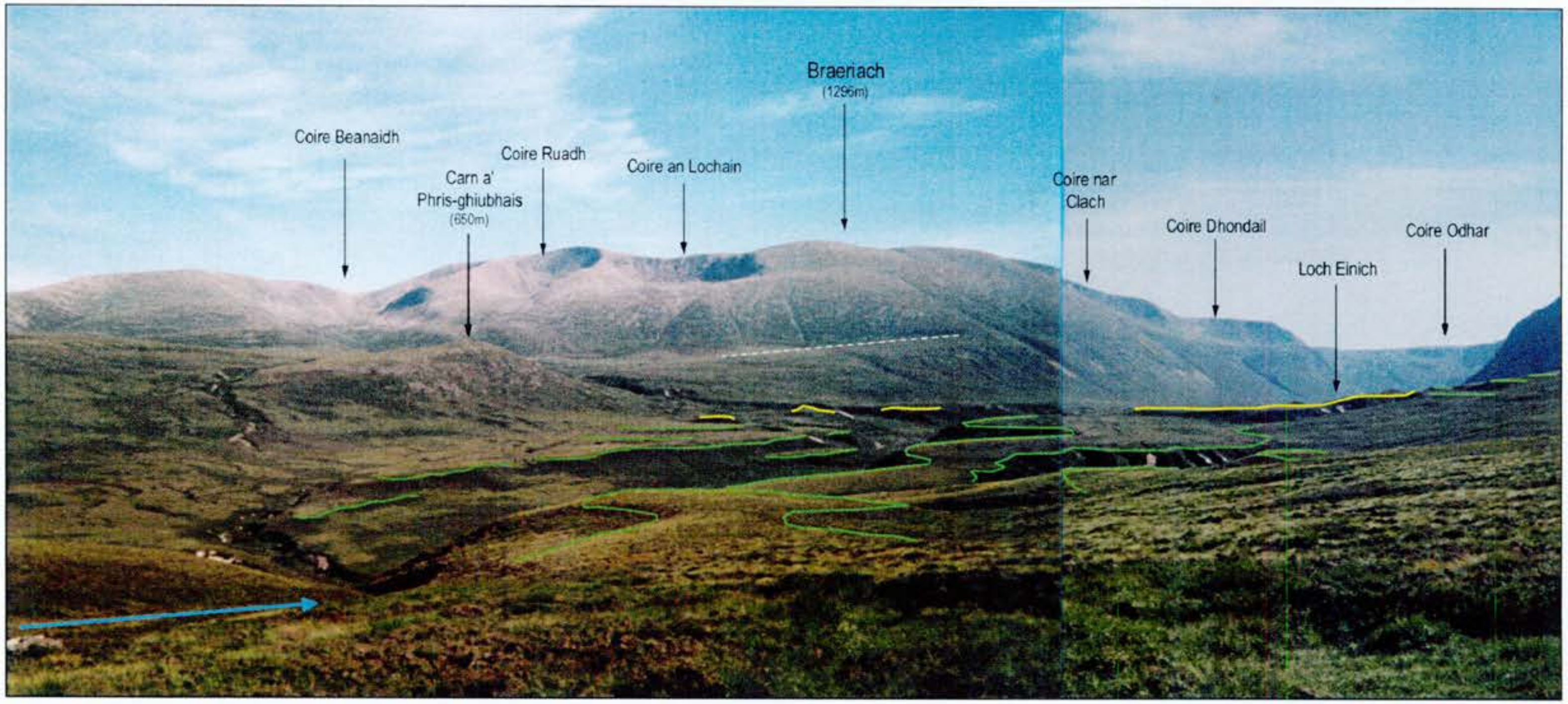


Figure 3.11: View southwards from the Glen More moraine, into Glen Einich. The extensive lake terrace and delta systems (green) can be seen on the valley floor, and in the immediate foreground. The eastern terraces appear on the left of the photo, whilst the deltas draining from Coire creagach are in the background on the right. The Glen Einich terminal moraine complex (yellow) crosses the floor of the glen to the south of the delta features. The meltwater channel draining from Glen More can be seen in the foreground (blue arrow). A series of drift trimlines are also visible on the flank of Braeriach (white dashed line).

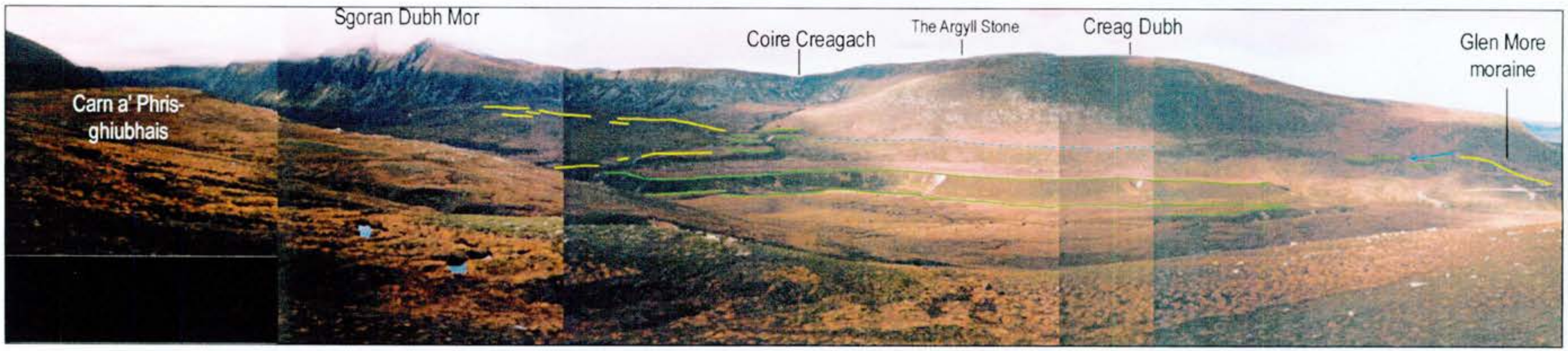


Figure 3.12: Glen Einich palaeo lake basin from the southern flank of Carn Eilrig. Lake shorelines (dashed blue line) associated with delta sets draining from Coire Creagach (green), and delta sets and meltwater channel (blue arrow) draining from the Glen More lobe, over the Glen More moraine (yellow) can be clearly seen on the eastern flank of Creag Dubh.

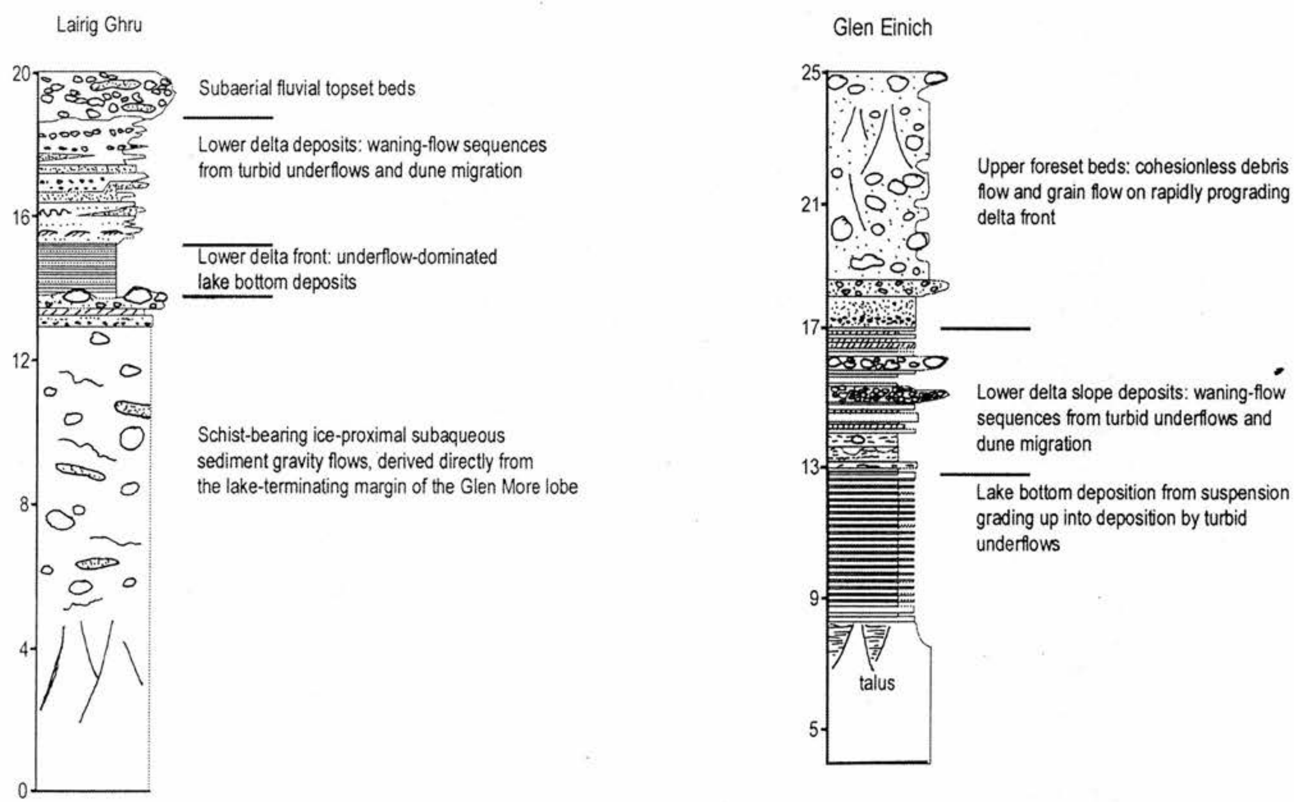


Figure 3.13: Lithological logs of sections in distal locations of the Lairig Ghru and Glen Einich deltas (after Brazier et al, 1998)

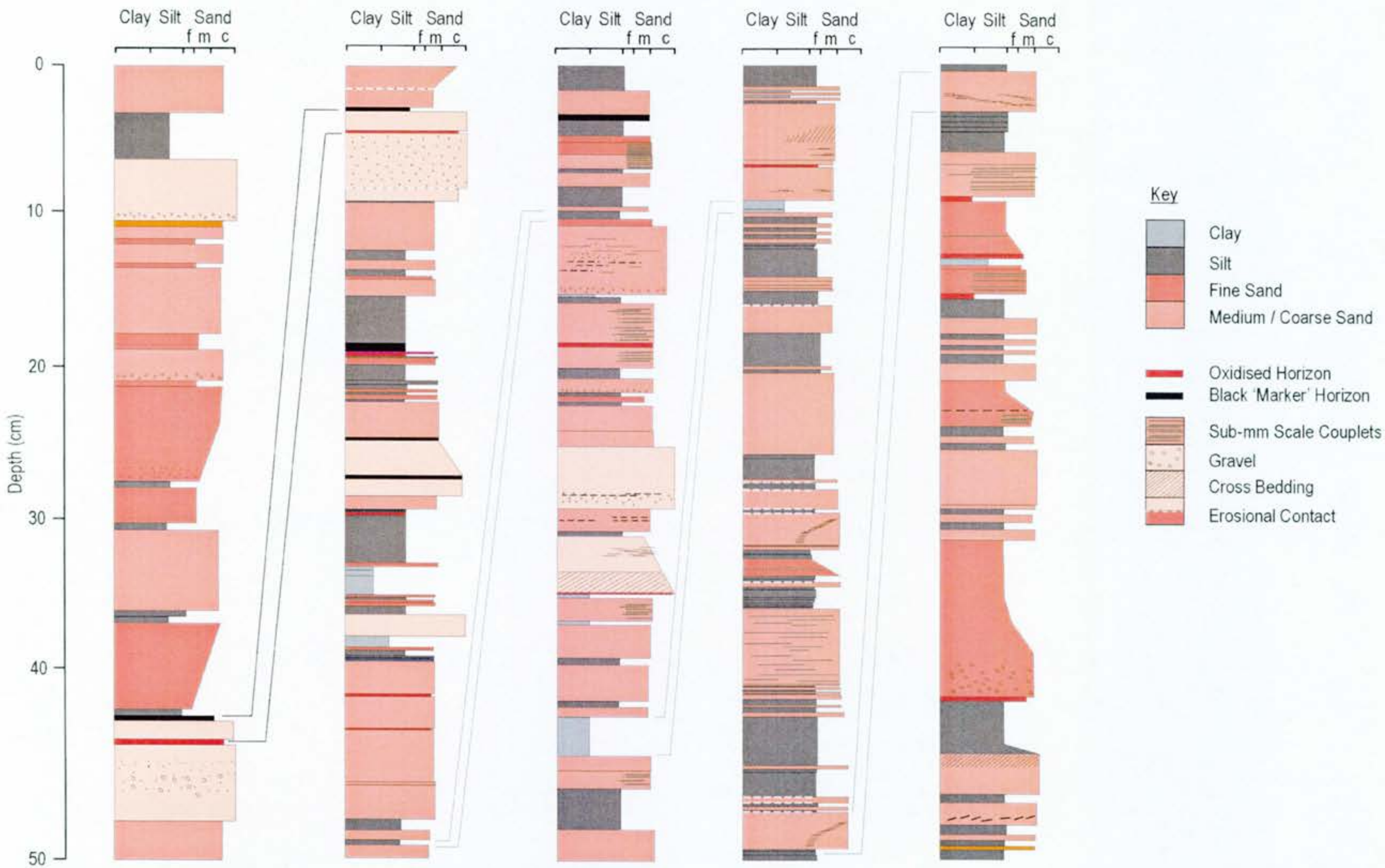


Figure 3.14: Example of a section of the Glen Einich valley floor delta sequence. The fine complexity of the sedimentation can be seen throughout the entire section.

Figure 3.15a: Exposed section of Glen Einich valley floor lacustrine delta sets.
Figure 3.15b: Close up of exposed section- lamination is evident at the right hand side of the figure, beneath which slumping obscures the internal structure.
Figure 3.15c: Cleaned section of exposure in figure 3.14. Clay/silt couplets are evident in the section, a pattern which repreats discontinuously for several metres vertically.
Figure 3.15d: Exposed section of delta sets in the Lairig Ghru, at the outflow of the meltwater channel draining from Glen Einich. Here the rhythmic silt/clay couplets are uncomformably draped over diamict, revealing abandonment of the site by ice, followed by submergence by an ice-dammed, proglacial lake.

3.3: Glen Geusachan

Glen Geusachan feeds into Glen Dee from the Braeriach/Moine Mhor plateau in the west, and it joins Glen Dee south of the Devils Point. To the east is a low col leading into Glen Luibeg. The glen closely resembles a U-shape, with steep sides and a flat floor (Figure 3.16 and 3.16a). The cliffed sides of Glen Geusachan rise steeply to elevations of 600m above the valley floor. The valley floor shows prominent channel and ridge features.

3.3.1: Landforms

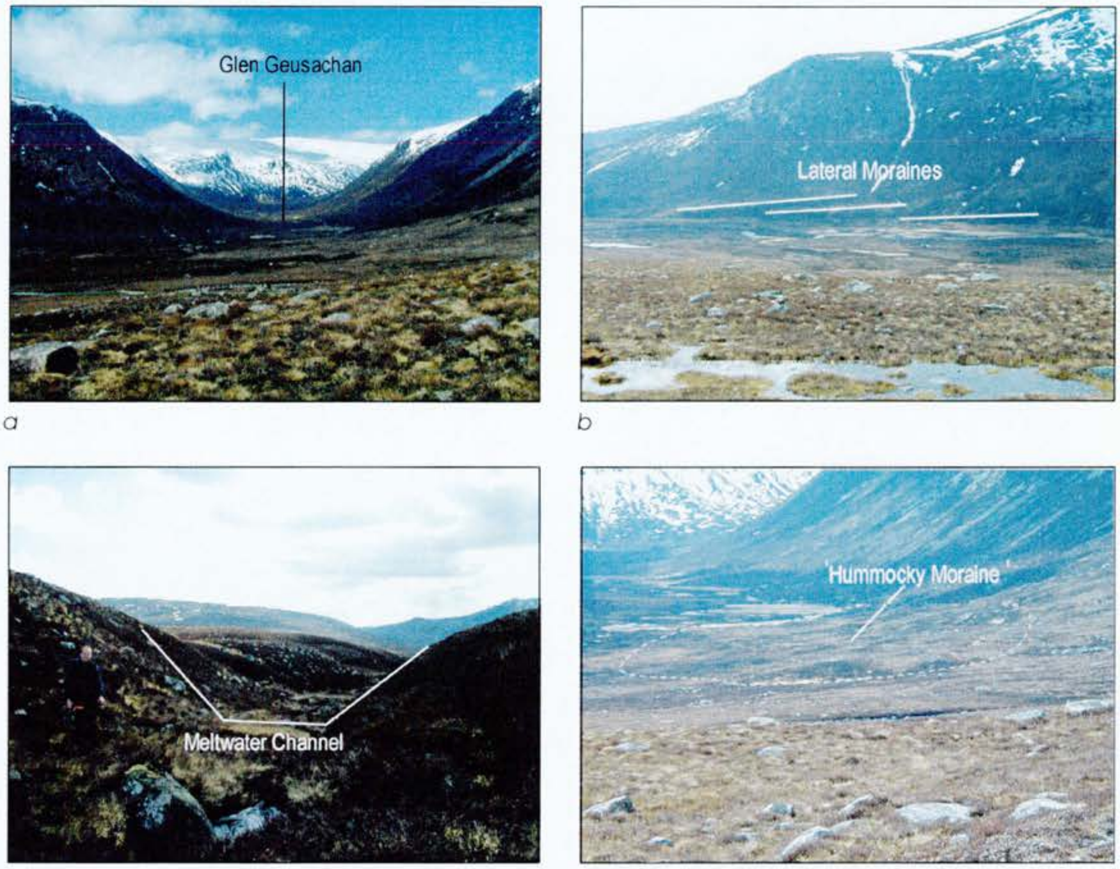
3.3.1a: Moraines

Most of the evidence for glacial activity within Glen Geusachan lies in Glen Dee. At the southern end of the glen there are a series of three ridges, clearly identifiable on the eastern side of the valley. Their extension westward is truncated by a breach partially caused by the channel of the present river Dee, but also by a system of relic channels. To the south of the ridges, predominantly on the western side of the Dee's present course are a series of large hummocks, unsorted boulder deposits, and channels covering a 1km^2 area. Within the ridges on the western side of the valley, a clear ridge can be seen, with its lowest limit at NN980933, trending upwards along the valley side at angles of around 3 - 5°, and finally disappearing at the junction between regolith and smoothed bedrock at the point where Glen Geusachan joins Glen Dee from the west (Figure 3.16b). The ridge is some 3 m in height, and exhibits a shallow, though short western slope, and a longer steeper eastern slope into the valley. There are lower ridges which are less well defined, but share similar slope angles and morphology to the upper ridge.

3.3.1b: Marginal moraine and meltwater features

On the eastern side of Glen Dee, directly to the east of the mouth of Glen Geusachan, a series of spectacular channels and associated moraine ridges can be seen. The moraine ridges between the channels display gentle proximal slopes, and steep distal slopes, though the modification that has occurred as a result of flow in the channels may have exacerbated this morphology. These moraine ridges clearly define a series of glacial limits in Glen Geusachan and Glen Dee (Sugden, 1970, Sissons, 1979, Bennett and Glasser, 1991). Moreover the absence of schist material within or on top of these moraines argues for the fact that the ice was locally sourced (Sugden, 1970). The outermost channel lies outside the easternmost moraine ridge, and is only a few metres lower than the elevation of the

Figure 3.16: Panoramic view of Glen Geusachan and Glen Dee, viewed from Meirlich Col (GR NN993937)



a b c d

Figures 3.16: Glen Geusachan Geomorphology:
a: Overall morphology of the 'U-shaped glen, leading into Glen Dee (foreground). Monadh Mhor and Braeriach in background. 'Hummocky moraine' on valley floor beneath Devils Point (right-midground).
b: Lateral moraines on the western side of Glen Dee, from the Geusachan Glacier. These are continuous with the terminal moraines in the southern part of the glen.
c: Meltwater channel on the eastern side of Glen Dee, formed at margin of Geusachan Glacier. Channel is one of a sequence of similar features, separated by moraine ridges. Channels are typically 3 - 4 m deep, flat bottomed, and dip to the south at angles of 3 - 5°.
d: Area of 'Hummocky Moraine', described by Sissons (1979) and Bennett & Glasser (1991). Originally thought to be remnants of Loch Lomond Stadial glaciation of Glen Geusachan.

col between Glen Dee and Glen Luibeg (Figure 3.16c and 3.17). This outer channel is intersected by a further channel, described by Sugden (1970) as ice-directed, indicating meltwater flowing eastward over Meirleach Col. Apart from smoothed bedrock features and a ridge high on the flanks of Creagan nan Gabhar, this is the only significant glacial geomorphological feature in the col. The channels on the eastern side of Glen Geusachan abutting the col, are up to 5m deep, and 10m wide, and are flat bottomed, and slope to the south at angles of 2 - 3° (Figure 3.16c).

3.3.2: Other evidence

The geomorphology of the northern extent of Glen Geusachan in the region of the Corrour Bothy is unclear. This area is covered with thick blanket peat, and few features can be confidently identified. To the north of this point the floor of Glen Dee is flat, and the valley sides display parallel horizontal features, indicated by a boundary between frost shattered upper slopes, and protected lower slopes. Further north in Glen Dee a ridge can be seen crossing the floor of the glen at NN976978. This ridge is a linear feature, truncated by the River Dee draining from Garbh Coire. To the south of this ridge, the valley sides within Glen Dee show clear horizontal transitions in vegetation type at 600 m. This change is associated with a break in slope, most easily noticeable in the southern part of Glen Dee, in the region of the Corrour Bothy, and has been associated with the presence of a palaeo lake (Bennett & Glasser, 1991).

3.3.2a: 'Hummocky moraine'

'Hummocky moraine' is a term that has been widely used to describe areas of disordered or chaotic moraine topography (Sissons, 1979, Boulton & Eyles, 1979, Eyles, 1979, 1982, 1983, 1993, Bennett, 1991, 1994, Bennett & Glasser, 1991, Bennett & Boulton, 1993, Anderson, 1998, Benn, 1992, 1993, Hambrey *et al.* 1997, Wilson & Evans, 2001). However the description is one that has become more generic in its definition, compared to the genetic classification of the landforms used by earlier authors (eg. Sissons, 1979). In its most basic use, 'hummocky moraine' describes moraine features that display no lineation or order in their plan form morphology. Typically the sedimentology of these features is characterised by interbedded debris flows and other mass movement deposits, laminated lacustrine sediments and glaciofluvial sands and gravels (Benn & Evans, 1997). Such features have been suggested to have been formed by stagnation processes (Eyles, 1979, 1982, 1983, Sissons, 1979 and others). More recent studies have highlighted distinctions between chaotic topography, and areas of hummocks displaying distinct lineation (Bennett & Glasser, 1991, Benn, 1992, 1993), arguing that formation is as a result of active ice retreat.

The floor of Glen Dee formerly occupied by the Glen Geusachan glacier is characterised in the northern area by a system of small ridges. These landforms extend from the foot of the Devils Point, southwards to around NN980940, covering an area of 1.5km^2 (Figure 3.16d and 3.17). Hinxman and Anderson (1915), Jamieson (1908) and Sissons, (1979) described these features as representative of an actively retreating ice mass, in contrast to Sugdens interpretation of them as remnants of widespread areal stagnation of a glacier downwasting in situ. Sissons assigned a Loch Lomond Stadial age to these moraines, in opposition to Sugden who suggested they were features formed during an earlier period of deglaciation of the post-LGM Cairngorm ice cap. Bennett and Glasser (1991) identified a bi-modal population of lineations within these features, and argued they were an active retreat feature, though did not place an age upon them. There are no ridges identifiable between the southern extent of these features, and the ridges to the southern end of Glen Dee. From above, the features in Glen Geusachan appear to be a discontinuous series of sub-parallel ridges (Figure 3.1). There is a distinct pattern of arcuate or lobate-fronted features, curving into Glen Dee from the northern end of Glen Geusachan.

To the west, within Glen Geusachan itself, the valley floor shows few landforms other than those formed from mass movements, such as talus slopes. High on the south western side of the glen, where the main body of Glen Geusachan turns from its upper southerly direction, to follow the eastward trending route of the main valley, Coire Cath nam Fionn does exhibit a series of small ridges at an altitude of 740 m.

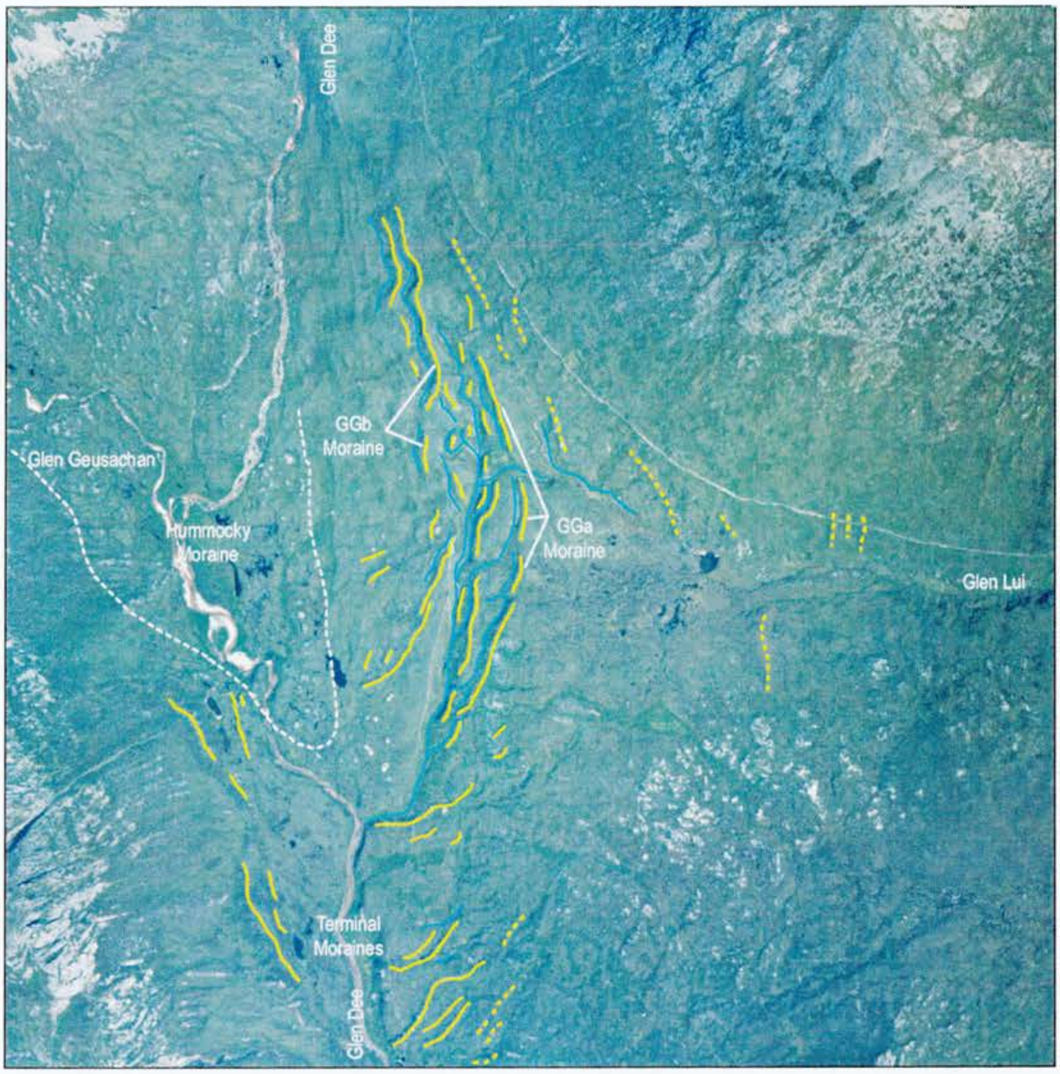


Figure 3.17: Colour aerial photo image of Glen Geusachan/ Glen Dee. Lateral moraine sequences (yellow) including the two sampled moraines (GGa and GGb) can be seen on the eastern side of Glen Dee. The outer terminal moraines at the southern end of the glen date from a slightly earlier phase of glaciation, where ice from Geusachan occupied the col separating it from Glen Luibeg. These moraines did not support boulders suitable for dating, and they also represent the period immediately prior to the stabilisation of the glacier within Glen Geusachan. As such it was felt that the stable stage of the glacier would provide more information to the study, rather than one of a suite of retreat morianes (yellow dashed).
Evidence that ice overtopped the col is provided by the meltwater channel draining from Glen Dee, over the col into Glen Lui. The area of 'hummocky moraine' represents the rapid retreat following the phase in deglaciation that created the moraine ridge and meltwater complexes. For more detailed mapping of the moraine sequences, see Map Insert.

3.4: Interpretation of Geomorphology

This section attempts to synthesise the geomorphological evidence into a reconstruction of activity in each of the three main areas of study. By examining the evidence gathered here, and by earlier researchers (Young, 1974, Sugden, 1970, Sissons, 1979, Bennett & Glasser, 1991, Brazier, *et al*, 1996, 1998) a relative chronology has been constructed for the western Cairngorms.

3.4.1: Decline of glaciation following the LGM

Throughout most of the deglacial phase at the end of the LGM, ice occupied the Glen More basin to a significant thickness. During much of this time, this ice from the Spey Valley, fed from the south west by the main Scottish Ice Sheet, coalesced with ice flowing out of the Cairngorm glens. A phase of westward retreat and thinning can be identified in the geomorphology, culminating in the final decoupling of the two glacial systems at the mouth of Glen Einich (Young, 1974, Sugden, 1968, 1970, Brazier et al, 1996, 1998, Golledge, 2002).

The highest stand that is identifiable on the northern flanks of the massif was identified by Sugden (1970) as the upper limit of schist deposition at an altitude of between 600 – 650 m on the northern flanks of the massif. There are no other easily identifiable limits above this in this part of the Cairngorms, save those in the corries themselves, which are argued to have been formed during the Loch Lomond Stadial (Sugden, 1970, Purves et al, 1999, Rapson, 1985, Gordon, 2001). If one accepts that this upper limit represents the margin of the lobe of ice in the Glen More basin, then it must be assumed that the basin was therefore filled by a significant lobe of ice from the main Spey Valley glacier. By simple calculation of surface gradient, using Nye's equation (1952), a lobe of this width across Glen More and the Spey Valley represents a thickness in the middle of the Spey of 420 m. Again a simple calculation of surface slope reveals that the snout of the Spey glacier must have been around 8 km to the north of the Glen More basin. At this time, ice draining the Lairig Ghru and Glen Einich would be coalescent with that in Glen More. Similarly ice from Strath Nethy would have joined with Spey Ice (Sugden, 1970).

At least three successively lower altitude phases may be clearly identified both in the field and from air photo imagery, the final phase being that associated with the moraine damming the mouth of Glen Einich (Figure 3.19). Within the lowest limits, a different pattern may be seen. The effects of meltwater are more obvious- esker systems, meltwater channels and outwash fans appear to dominate the floor of Glen More, with the central feature being the kettle hole formed by Loch Morlich (Young, 1974).

3.4.2: Deglaciation of Glen Einich

3.4.2a: Early deglaciation of Glen Einich

There is ample evidence for Glen Einich and the surrounding hills to have been completely overridden by ice during glacial maximum conditions (Sugden, 1968, Rea, 1991, Brazier et al, 1998).

To the south of the headwall of Glen Einich the Moine Mhor plateau displays little direct evidence of ice coverage, as blanket peat shrouds the underlying topography. However this plateau surface is a low-lying area, surrounded on all sides by once ice-covered hills. It also must have provided the source area for the majority of the ice flowing in Glen Einich, additional input draining from Coire Bogha-cloiche, Coire Dhondail and Coire nan Clach, on the flanks of Braeriach.

As thinning of the Glen More lobe continued, it was accompanied by a retreat of the Glen Einich glacier, which was lowered from its eastern marginal position on the col between Glen Einich and the Lairig Ghru. The channel draining across the col between Glen Einich and the Lairig Ghru has been interpreted as resulting from meltwater draining a high stand of the Glen Einich glacier, at a time when the Lairig Ghru glacier was in retreat (Brazier et al, 1998). The altitude of the channel implies that ice in Glen Einich was probably occupying areas of the col between the two glens, and thus was in coalescence with ice in Glen More. The Lairig Ghru glacier had clearly separated from the Glen More lobe at this stage, evidenced by the extensive ice dammed lake between the two ice masses. Thus this represents firm evidence for separate ice limits of both regional and local ice.

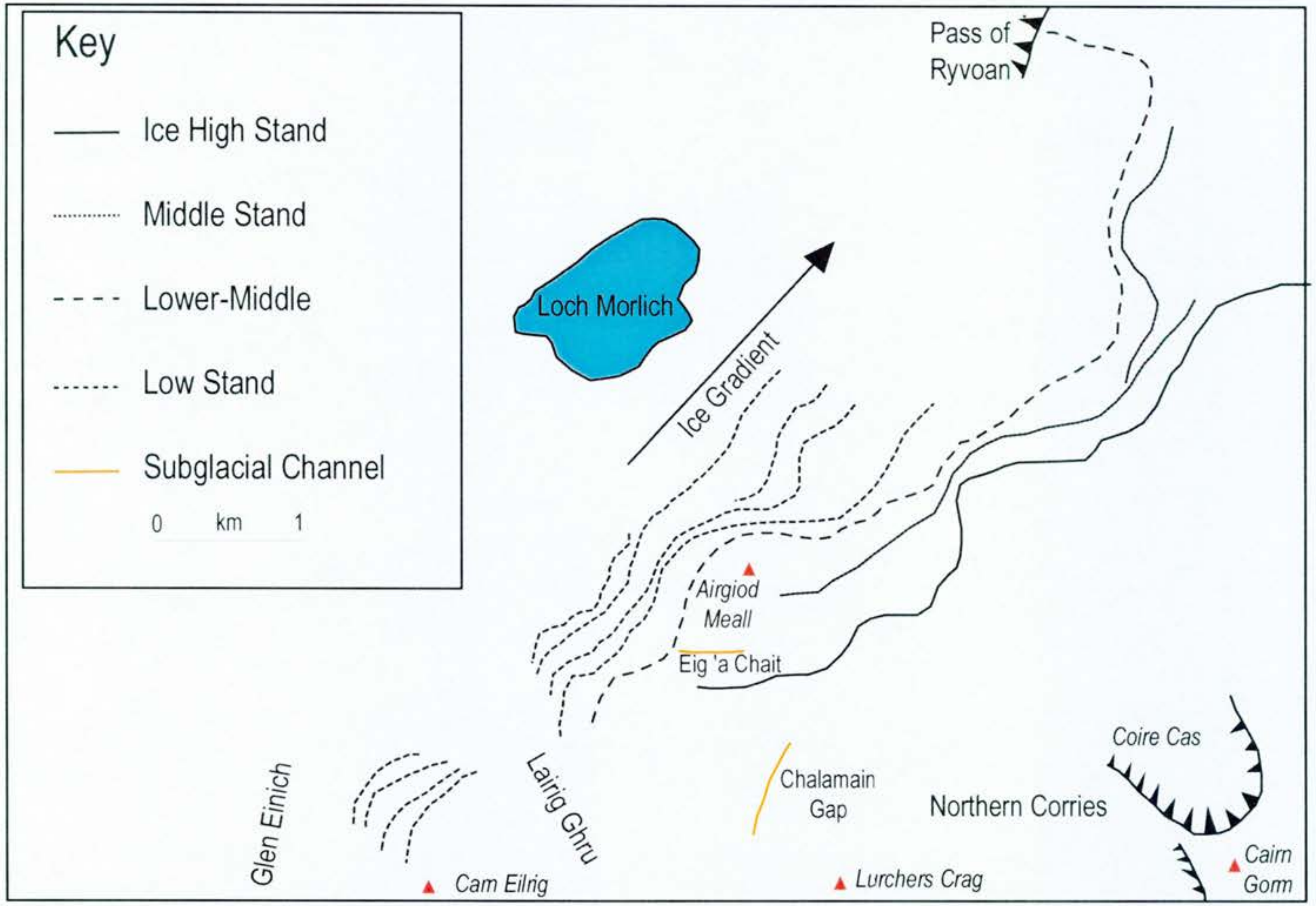


Figure 3.19: Reconstruction of Glen More lobe glacial limits on the northern Cairngorm flanks. Several stands can be identified, associated with meltwater channels.

3.4.2b: Separation of the Glen Einich glacier and the Glen More lobe

The Glen More lobe, being supplied by ice from the main Scottish Ice Sheet, maintained a relatively stable snout position within the mouth of Glen Einich, most likely pinned by the flanks of Carn Eilrig and Cadha Mor (Figure 3.20). As a result it formed an effective ice dam to meltwater attempting to flow downslope to the north, both from the snout of the Glen More lobe, and from the Glen Einich glacier , thus creating an ice dammed lake on the floor of Glen Einich (Figure 3.21). Situations analogous to this can be seen at Gokyo Ri in the Himalaya, where the Ngozumpa glacier has dammed a series of lakes in tributary valleys (Figure 3.22), and on Axel Heiberg Island in Arctic Canada, where the Thompson Glacier has dammed an extensive series of lakes, the largest being around 5 km long (Hambrey, 1994) (Figure 3.23).

3.4.2c: The Glen Einich ice dammed lake

The Glen Einich lake was in existence later during deglaciation than the lake in the Lairig Ghru, and the sedimentological evidence argues for its existence over several centuries (Brazier *et al*, 1998). Lake shorelines on the distal side of the Glen More moraine illustrate two low positions of the lake identified by Brazier, however shorelines on the flanks of Carn Eilrig (520 and 510 m), and also on the eastern slope of Creag Dubh (500 and 480 m) that have not previously been identified, indicate that the highest stand of the lake was at least at the height of the top of the Glen More moraine. Further evidence for this high stand is to be found in the deltaic sequences, standing at around 520 m beneath Coire Creagach and at the upper level of the western side of the Glen More moraine (Figure 3.20). It is not known how long the lake occupied this high stand position. However the limited extent of deltaic features at these two locations, in comparison to the much more extensive lower features on the valley floor, argue for only a short period of high stand.

Examination of the stratigraphic sequence within the largest delta in Glen Einich reveals an outwardly simple story. Brazier argues that the sequence shows a classic deltaic succession from rhythmic sedimentation in a distal lake floor environment, to increasingly proximal lower and upper foreset deposition, as the delta front prograded southwards down the lake (Brazier *et al*, 1998). The predominance of clay and silt couplets in the lower layers (Figure 3.13, 3.14 and 3.15a, b, c and d), below 13-14 m possibly indicates low energy sedimentation in a seasonally frozen lake. The presence of dropstones indicates active calving into the lake, though their rarity argues that glacier snouts were not water-terminating for long during the lakes existence. A second line of evidence suggesting that snouts were actively calving into the lake is the plan form shape of the Glen Einich terminal moraines. These are concave in form up-valley, similar to those described for calving snouts in the Western Highlands (Benn, 1996) and Patagonia (Warren and Sugden, 1993). Brazier argues that the Glen Einich lake was dammed by the Glen More moraine and ice lobe to the north (Figure 3.24), until the lobe thinned sufficiently to allow drainage subglacially, in combination with continued drainage along the ice margins. Throughout this time the lake appears to have been stable.

In summary therefore, the landforms of Glen Einich, Glen More and the northern flanks of the Cairngorms document the thinning and retreat of the Scottish Ice Sheet from the north west. The main inferences to be taken from this interpretation are:

1. The northern flanks of the Cairngorms saw a successive thinning of the Scottish Ice Sheet in Glen More. This thinning took place during a series of phases, punctuated by halts or stillstands in retreat (Figure 3.19).
2. A stillstand of both local and regional ice took place indicated by the Spey Ice moraine at the mouth of Glen Einich. The timing of this stillstand is known to be coincident in both the Cairngorm glaciers and Scottish Ice Sheet, by interpretation of lake deposits in Glen Einich (Brazier, *et al.* 1996, 1998).

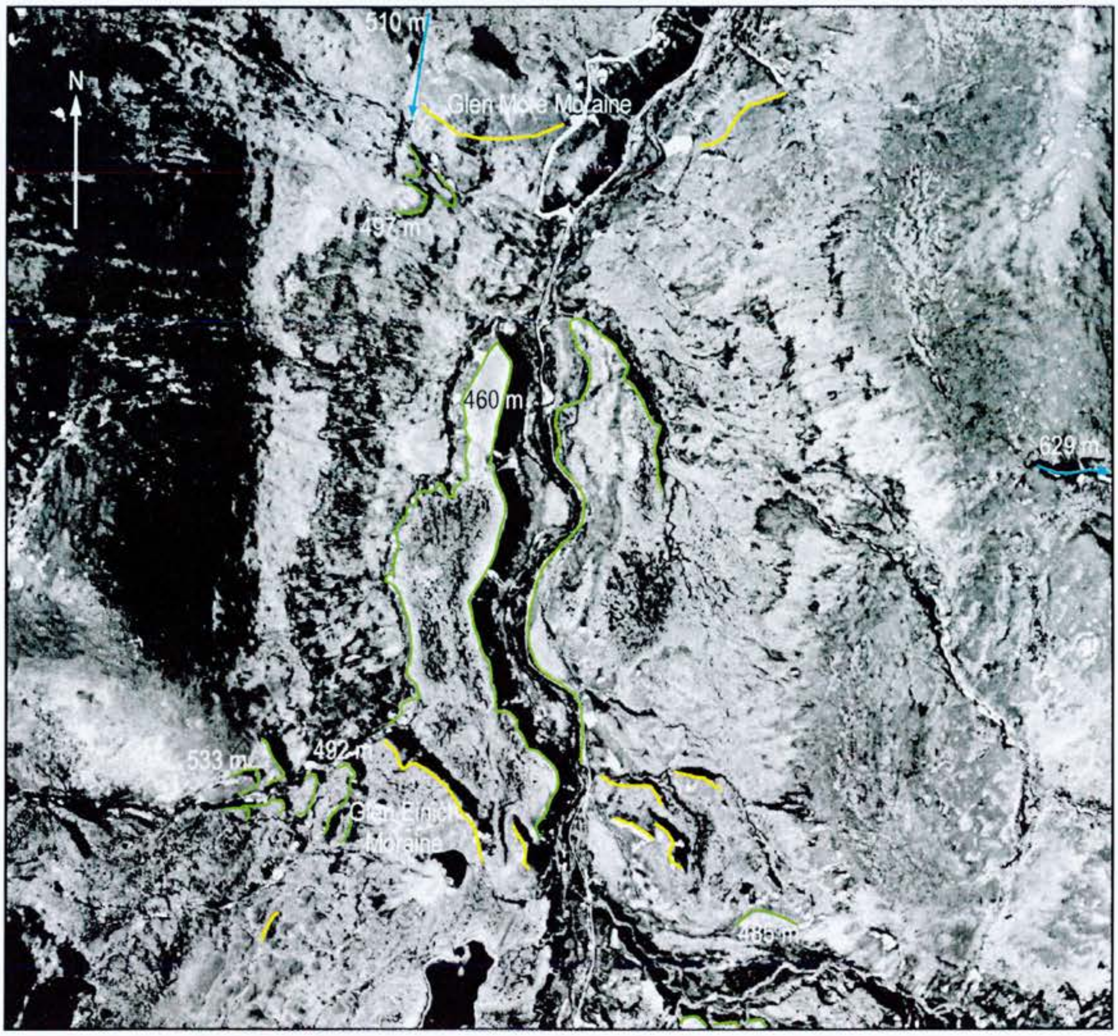


Figure 3.20: The lake was dammed to the north by Glen More ice, and to the south by the Glen Einich glacier. Ice calved into the lake for at least some of its lifetime. Delta sequences surrounding the lake indicate show drainage from a high stand of at least 70 m above the valley-floor deltas. A meltwater channel drains into the lake over the Glen More moraine, and associated delta sets beneath show that this occurred during a lake stand of 30 m above the valley floor deltas. A meltwater channel drains from Glen Einich over the col into the Lairig Ghru at 629 m, evidence of an earlier higher ice margin in Glen Einich.

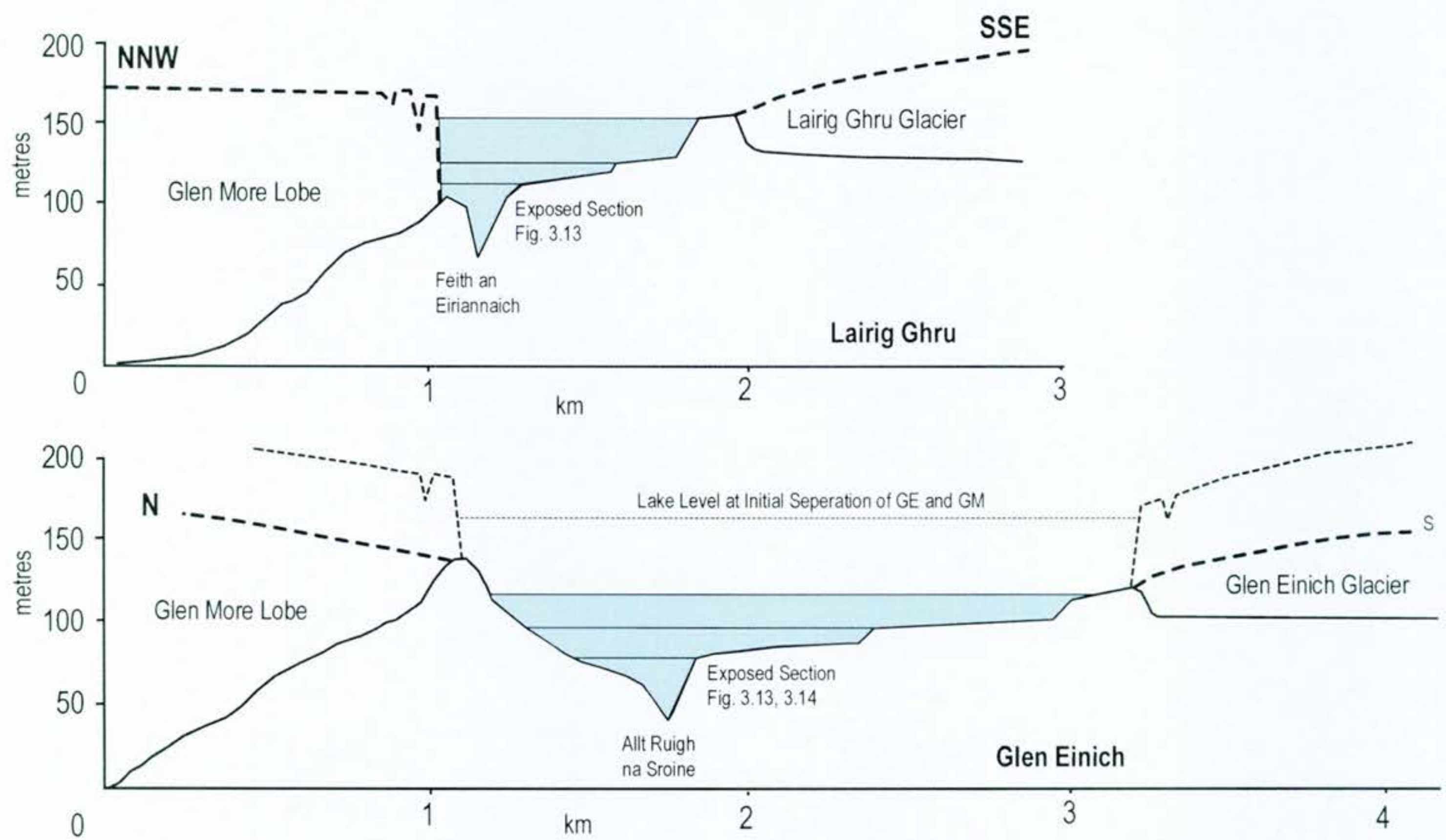


Figure 3.21: Diagrammatic reconstruction of the Lairig Ghru and Glen Einich ice dammed lake systems (after Brazier et al, 1998)

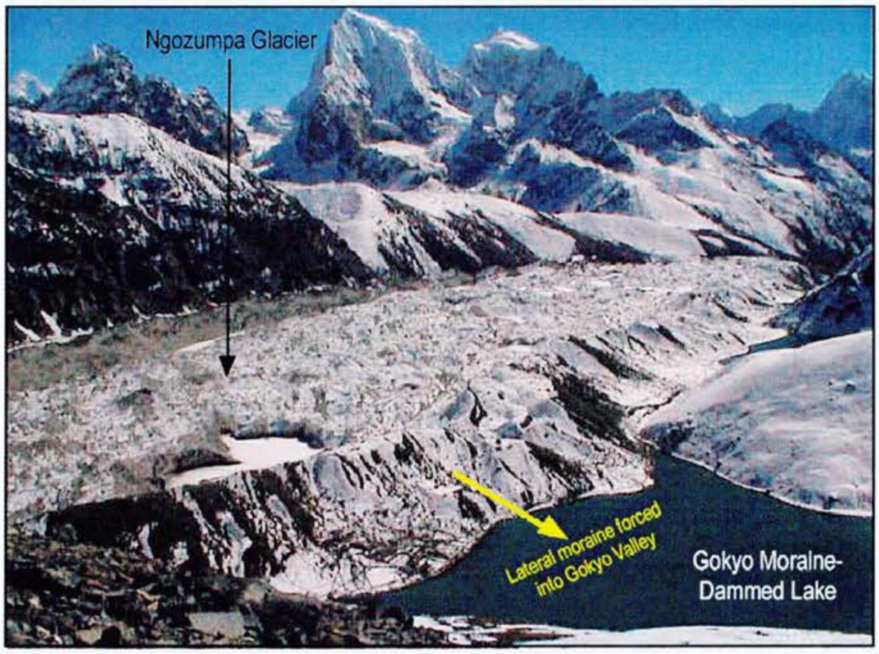


Figure 3.22: Gokyo from Gokyo Ri, Nepal. A modern analogue to the situation in Glen Einich during late stages of deglaciation. The Ngozumpa has built a large lateral moraine ridge into the Gokyo Valley, thus damming a lake, fed by meltwater. An earlier, higher stand of the ice may have created an ice dam, exactly analogous to Glen Einich.

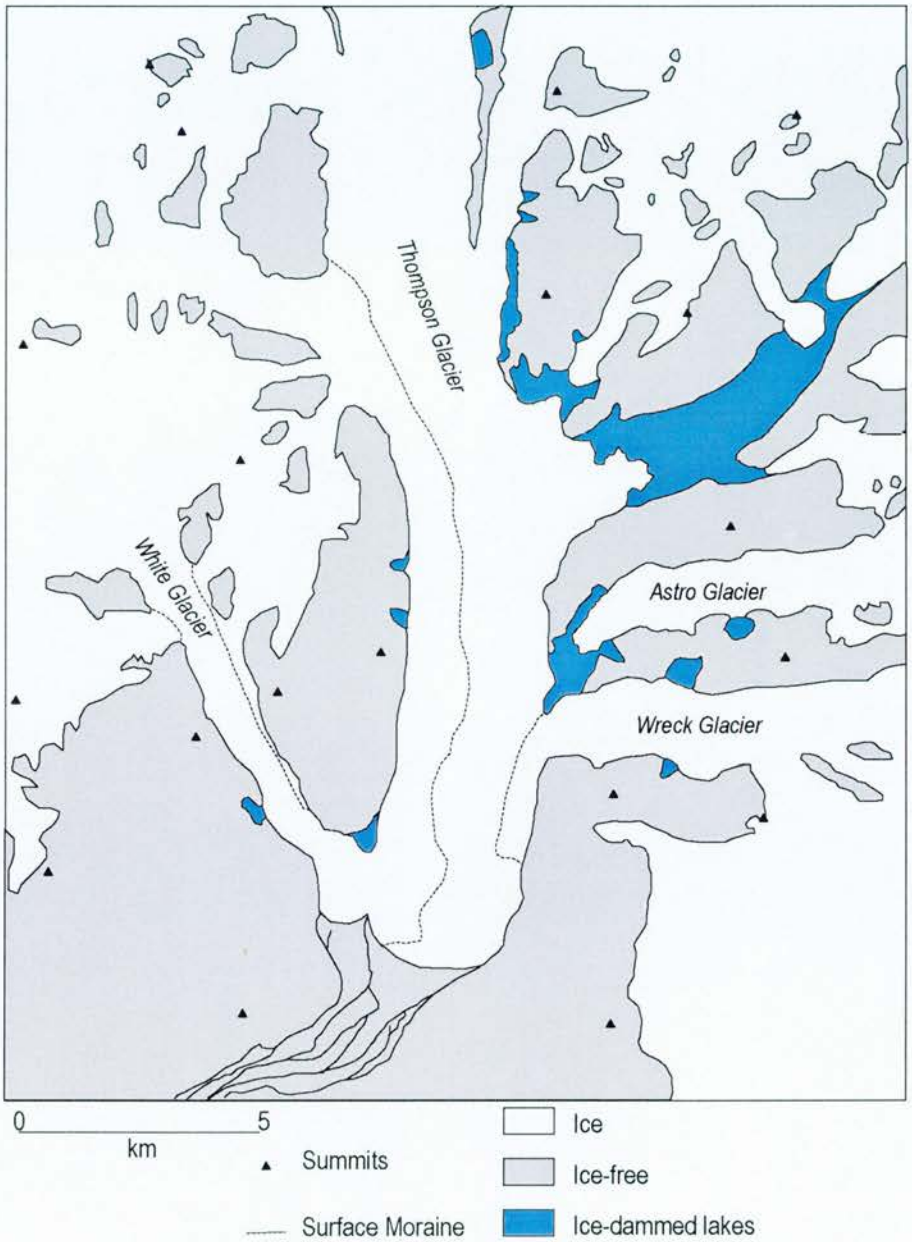


Figure 3.23: Ice-dammed lakes associated with the Thompson Glacier, Axel Heiberg Island in Arctic Canada (Hambrey, 1994)

3. The lake deposits show that the lake was impounded in the Glen by ice in Glen More, and was actively fed by meltwater and bergs from ice in Glen Einich, draining from the local ice cap. The deposits also show that the lake was in existence for around 1 ka (Brazier, *et al.* 1996, 1998).

3.4.3: Deglaciation of Glen Geusachan

The rate and character of retreat of ice over the col separating Glen Geusachan from Glen Lui is difficult to reconstruct, as little geomorphology survives, some evidence potentially being buried beneath the blanket peat that exists on the col. It is likely that this retreat was rapid.

The final phase of deglaciation in this part of the Cairngorms is therefore represented by the glacier limits found in the Glen Dee/Glen Geusachan area (Figure 3.17). The moraine and meltwater channel features on the eastern side of Glen Geusachan / Glen Dee represent the final stable marginal position of a glacier draining from the Cairngorm summit plateau, down into Glen Geusachan.

This position is likely to correspond to limits in Glen Einich, The reasons for this are several. The first is the sheer size of the Glen Geusachan glacier. Even using Sissons estimation, which relies on a limited accumulation area separate from other glacier source areas around the Breariach / Cairn Toul massif, then the area of the Glen Geusachan glacier is 9 km^2. If one assumes that the glacier was fed by a single ice cap on the Moine Mhor / Braeriach plateau, then the approximate total size of the glacier is 13.5 km^2. The Glen Einich glacier, at its still stand position, has an approximate area of 9 - 12 km^2. The second line of evidence derives from these size relationships. The Equilibrium Line Altitude proposed by Sissons for his Glen Geusachan glacier is 730 m. Reconstructing ELA's for glaciers of similar size in the western Cairngorms reveals only two valleys have similar figures- Glen Einich and Garbh Coire (Figure 6.2). The implication being that these glaciers existed at the same time, fed by the same plateau icecap.

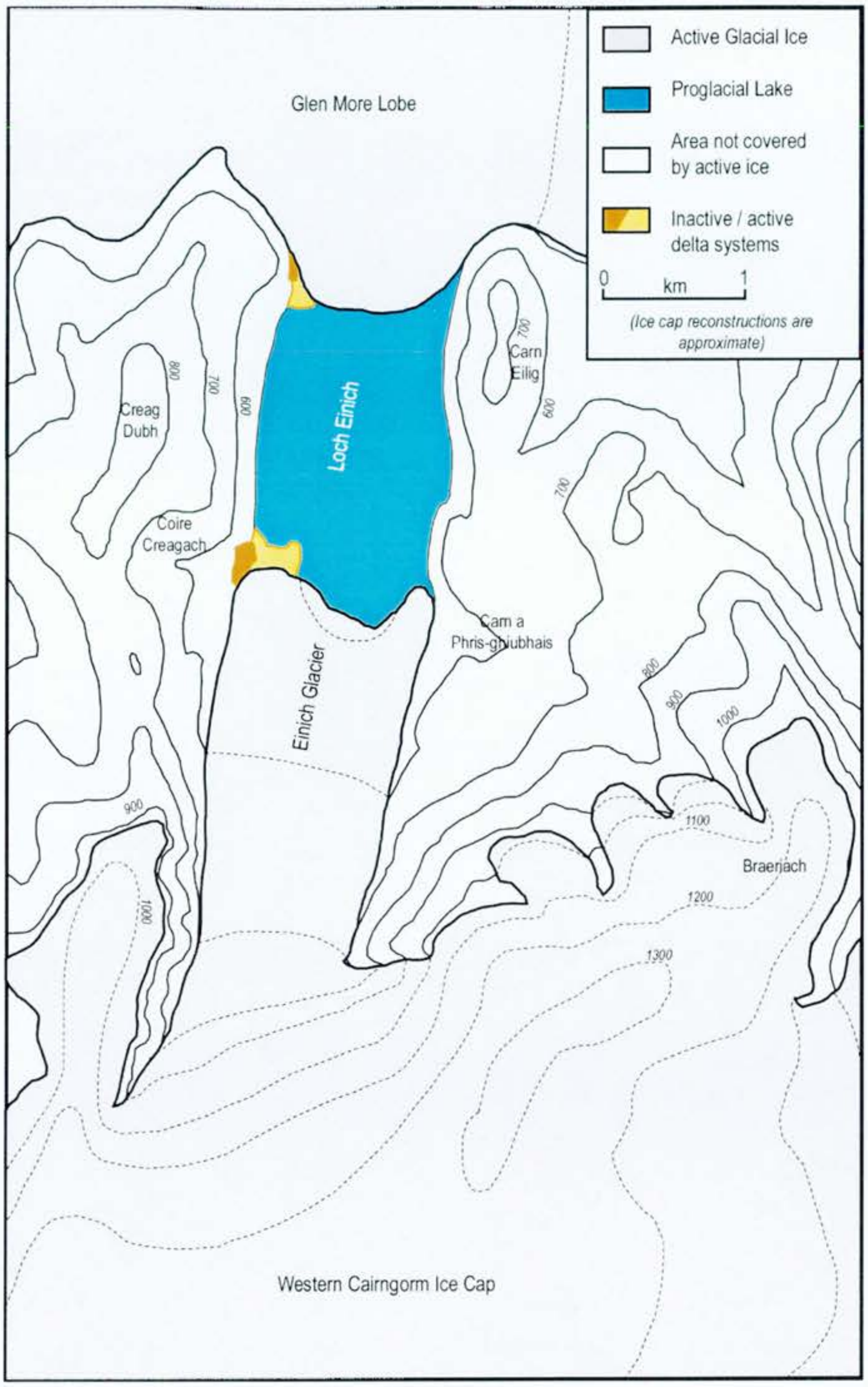


Figure 3.24: Generalised reconstruction of the glacial situation in Glen Einich, represented by the geomorphology described in this chapter. A large proglacial lake in Glen Einich is dammed to the north by ice in Glen More. The Einich glacier, an outlet of the small Western Cairngorm Ice Cap, calves into the lake. This situation persisted for around 1 ka. Ice cap limits are defined by moraine features on the apron to the north of the Braeriach corries.

A clear sequence of terminal ridges mark the Glen Geusachan glaciers southernmost extent, though its northern limit is less clear. From earlier workers and the limited geomorphological evidence, a limit has been placed spanning Glen Dee in the region around the Corrour Bothy, the overall shape of the glacier within Glen Dee can be described as 'piedmont' style (Bennett & Glasser, 1991). Lake shorelines mark the position of a shallow lake to the north of this limit, presumably dammed to the south by the Glen Geusachan glacier (Bennett & Glasser, 1991). Outflow from this lake flowed around the eastern margin of the glacier, and emptied into Glen Dee at the southern glacier terminus.

Following a period of relative stability at these margins, the Glen Geusachan glacier is argued to have retreated rapidly westwards. There are no clear classic terminal or lateral moraine positions within the dated limits, however the presence of 'hummocky moraine' beneath the southern flanks of the Devils Point have been argued to represent Loch Lomond Stadial glacier positions (Sissons, 1979, Bennett and Glasser, 1991). The genesis of the 'hummocky moraine' in Glen Geusachan is of secondary importance to its age. All authors who have worked in the Glen have agreed that the sequence of 'hummocky moraine' forms part of the same sequence of landforms resulting from the same glacial phase as that which formed the moraines dated in this study.

3.5: Summary

The limits identified in this section, and the glacial phases that they represent all form part of a relative chronology of ice sheet activity in and around the Cairngorms during the final phases of deglaciation at the end of the LGM. The key findings in the geomorphology of the western Cairngorms are:

1. The retreat of the Scottish Ice Sheet took place by progressive thinning and phases of retreat to the south west.
2. Potentially the last of these phases is represented by a stillstand damming the northern mouth of Glen Einich.
3. This stillstand can be correlated with glacier limits in Glen Einich formed by local, Cairngorm ice, as shown by a lake dammed between the two ice masses.
4. The lake in Glen Einich survived for at least a number of centuries.
5. The glacier in Glen Geusachan is similar in size and former ELA with that in Glen Einich. The two are thought to be contemporaneous, and representative of the

final stage of deglaciation of the Cairngorms by both local and regional ice systems.

Samples for cosmogenic surface exposure dating were collected from the Glen More moraine at the mouth of Glen Einich, the Glen Einich lateral moraines, and the Glen Geusachan lateral moraine complexes. These features represented the final stage of deglaciation in the western Cairngorms, indicated by the geomorphology. The dates collected therefore provide an empirical test of the interpretation of the geomorphology. They also provide an age for the disappearance of ice in the western Cairngorms, and from the Glen More basin. Additionally they provide resolution to a long standing debate over the extent and timing of valley glaciation in the Cairngorms during the Loch Lomond Stadial.

Chapter 4:

Chronology

Chapter 4: Chronology

This chapter discusses the results of both the cosmogenic ^{10}Be work from the Cairngorms and ^{14}C data from the Loch Etteridge core.

4.1: Cosmogenic Dating

Fifteen samples were collected and processed from three main locations; the Glen More moraine built by regional Scottish ice at the mouth of Glen Einich, and the marginal moraines created by local Cairngorm ice on the western side of Glen Einich, and finally the marginal moraines in Glen Geusachan.

4.1.1: Glen More Samples

The aim of sampling at these locations was to ascertain an age for the last occupation of the moraine ridge by Glen More ice. The size of the Glen More moraine required that sample selection was made with care. Few boulders on the ridge crest overlooking Glen Einich appeared to be suitable, since the construction of the bulldozed track may have disturbed some boulders. Furthermore, on the upper slopes of the western portion of the moraine, there are few large boulders. Therefore only two boulders were sampled from the crest of the moraine (Figure 4.1 and Table 4.1). The samples (CE1 and CE2) were both taken from boulders of 1.2 m height, located 500 m from the edge of the Rothiemurchus Forest. There was no evidence of local shielding, but it is likely that the moraine was forested throughout a large part of the Holocene, as shown by extensive pine stumps further south in Glen Einich. The second pair of Glen More samples (CE3 and CE4), were taken from boulders 1.8 and 1.5 m high respectively, on the proximal slope of the moraine, at the edge of the modern Rothiemurchus Forest. The density of tree cover at this location is currently low, with approximately one Scots Pine per 100 m^2, but earlier in the Holocene, cover would have been higher. All samples from the Glen More moraine experience low environmental shielding, as summits in this part of the Cairngorms are relatively low and distant.

4.1.2: Glen Einich lateral moraines

Sampling from these two ridges sought to determine an age for the most extensive stand of the Glen Einich glacier at the time a maximum-stand lake was impounded in lower Glen Einich. The lateral moraines of the Glen Einich glacier form clear ridges supporting large boulders (Figure 4.2 and Table 4.2). No suitable boulders existed on

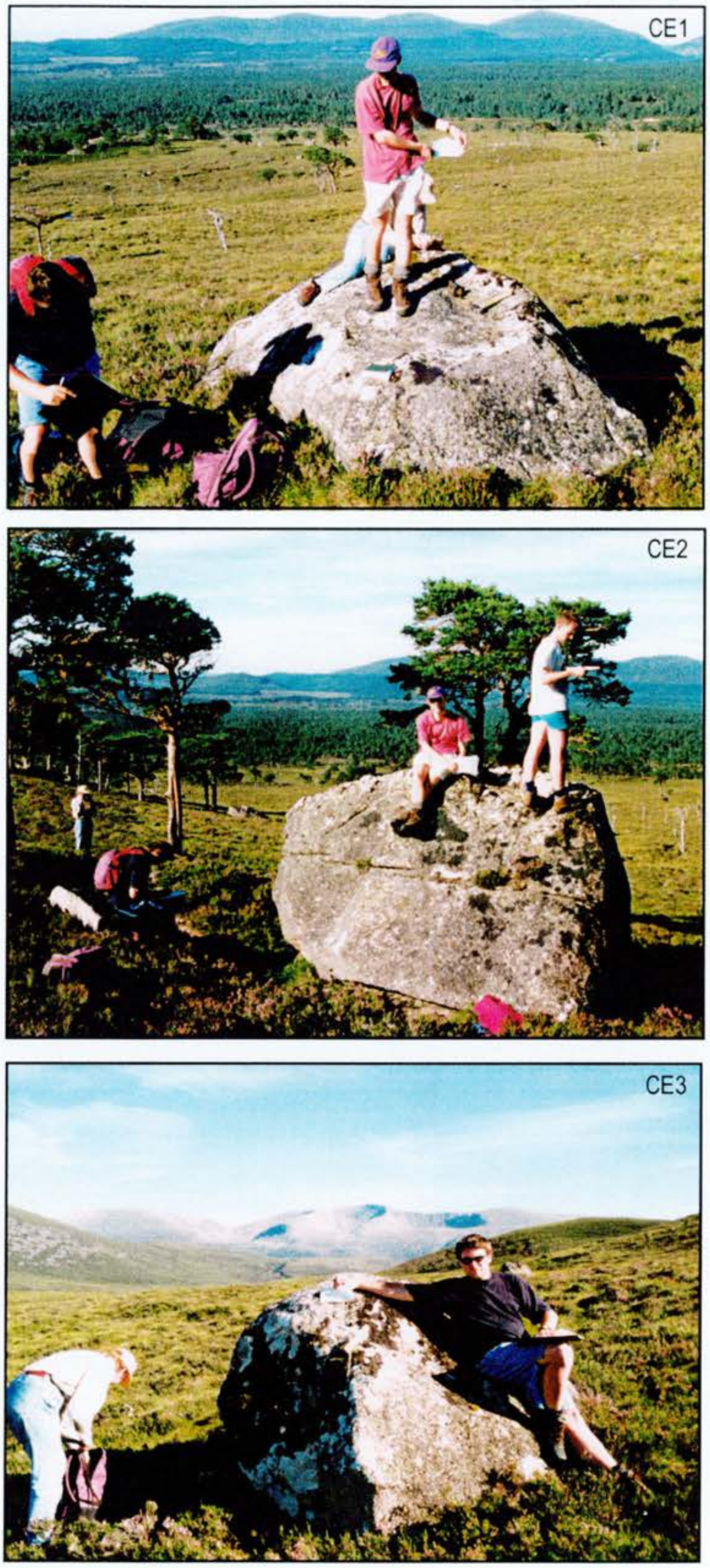


Figure 4.1: Boulders sampled from the Glen More moraine. CE3 and CE4 were the closest boulders to the ridge crest. CE1 and CE2 were on the proximal slope of the moraine, within an area of sparse tree cover. The sampling team included Dr Susan Ivy-Ochs, Dr Mike Bentley, Dr Ross Purves and Professor David Sugden.

Glen More Moraine

Sample	Location	Altitude	Lithology	Surface	Height	Context	Constraint Provided
CE1	57° 07' 49" N 3° 47' 2" W	0.43 km	Granite	Sub-angular, surface roughening. Some evidence of postglacial weathering	1.2 m	Proximal slope of Glen More moraine, on fluvioglacial ridge. 8 degree slope. Partial forest cover.	Retreat of Glen More ice from maximum position in Glen Einich.
CE2	57° 07' 56" N 3° 46' 42" W	0.48	Granite	Sub-angular, surface roughening. Some evidence of postglacial weathering	1.8	Proximal slope of Glen More moraine, on fluvioglacial ridge. 8 degree slope. Partial forest cover.	Retreat of Glen More ice from maximum position in Glen Einich.
CE3	57° 07' 41"N 3° 46' 36" W	0.47	Granite	Sub-angular, surface roughening. Some evidence of postglacial weathering	1.2	Crest of Glen More moraine, on 4 degree northward slope. Open outlook	Maximum extent of Glen More ice into Glen Einich, following separation of Cairngorm and regional ice.
CE4	57° 07' 49" N 3° 47' 5" W	0.41	Granite	Sub-angular, surface roughening. Some evidence of postglacial weathering	1.5	Crest of Glen More moraine, on 4 degree northward slope. Open outlook	Maximum extent of Glen More ice into Glen Einich, following separation of Cairngorm and regional ice.

Table 4.1: Glen More moraine boulder descriptions

the terminal moraine ridge crossing the valley floor. CE8 and CE9 were sampled from the outermost lateral moraine ridge, which runs north - south to the east of Lochan Beanaidh. This ridge is continuous for almost a kilometre, and displays a classic lateral moraine profile, being a broad ridge with steep distal and shallow proximal slopes. The ridge itself is up to 3 m high, and the sampled boulders were 1.7 and 3.5 m high respectively. The height of these boulders means that earlier Holocene tree cover will have had less of a shielding effect on these samples than those on the Glen More moraine. Likewise the size of these boulders mitigates against the possibility of significant post depositional movement. Environmental shielding here was the highest of all sample locations, due to the proximity of the cliffs south of Coire Creagach. Samples CE6 and CE7 were taken from a lower marginal moraine, running parallel to the outer moraine, 200 m to the east. This feature was less continuous than the outermost moraine, exhibiting a lower crest of roughly 1.5 m above the valley floor. Boulders on this moraine were also smaller, those sampled being 1.5 and 1.3 m high.

4.1.3: Glen Geusachan

The samples were collected from limits identified by previous authors (Sissons, 1979, Sugden, 1968, 1970, Bennett and Glasser, 1991, Brazier *et al.* 1996, 1998) as showing the extent of the independent Glen Geusachan glacier following ice sheet deglaciation. Samples collected from this location provide an age constraint for the final phase of deglaciation of Glen Geusachan, prior to the rapid retreat and collapse of the system at the end of the glacial period. Bennett and Glasser (1991, p. 119) argue that the associated channels demonstrate that the Glen Geusachan glacier occupied the position for a considerable period of time, since one of the channels is cut directly into bedrock.

4.1.3a: Outer ridge

Four samples were collected from the outermost moraine ridge on the slopes of Meirleach Col to determine an age estimate for the maximum extent of the Glen Geusachan glacier (Figure 4.3 and table 4.3). Unfortunately GGa3 was lost during sample processing, and was not sent to the AMS for dating. This 600 m long ridge, running from NN991939 to NN991933, is associated with a meltwater channel that runs continuously along its eastern flank. GGa1, GGa2, GGa3 and GGa4 were sampled from boulders of 1.5, 1.5, 1.2 and 1m height, all standing on the crest of the ridge. None of the boulders showed significant evidence of movement, such as

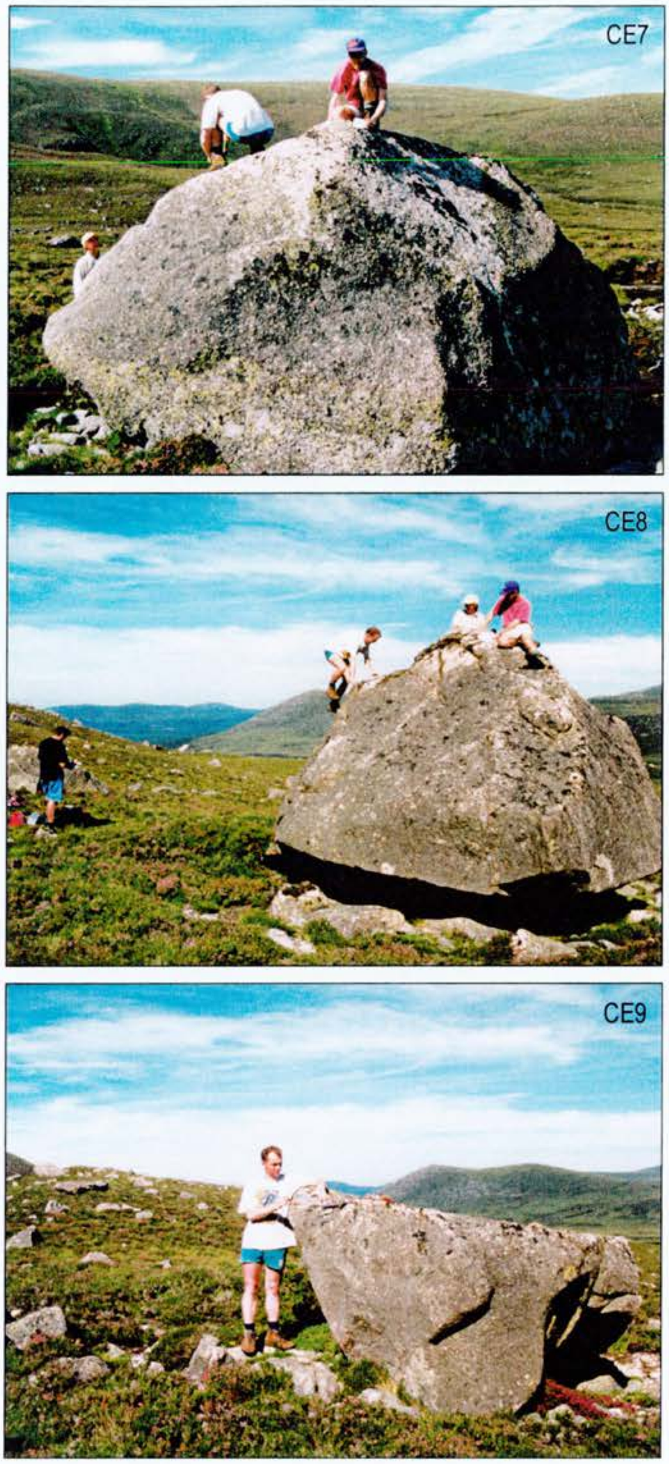


Figure 4.2: Boulders sampled from the Glen Einich lateral moraines. CE7 on the outermost moraine ridge can be clearly seen to have higher environmental shielding values than the other samples, due to its proximity to the cliff in Coire Creagach. All samples were thought to be large enough to escape significant winter snow coverage.

Glen Einich laterals

Sample	Location	Altitude	Lithology	Surface	Height	Context	Constraint Provided
CE6	57° 06' 23" N 3° 47' 59" W	0.5	Granite	Sub-angular, surface roughening. Some evidence of postglacial weathering	1.7 m	Crest of outermost Glen Einich lateral moraine, beneath Coire Creagach	Maximum extent of Glen Einich ice, following separation of Cairngorm and regional ice sheets. Contemporaneous with calving margin into lake in lower glen.
CE7	57° 06' 12" N 3° 47' 59" W	0.5	Granite	Sub-angular, surface roughening. Some evidence of postglacial weathering	2	Crest of outermost Glen Einich lateral moraine, beneath Coire Creagach	Maximum extent of Glen Einich ice, following separation of Cairngorm and regional ice sheets. Contemporaneous with calving margin into lake in lower glen.
CE8	57° 06' 2" N 3° 48' 6" W	0.52	Granite	Sub-angular, surface roughening. Some evidence of postglacial weathering	3.5	Crest of intermediate recessional Glen Einich lateral moraine	Thinning of Glen Einich ice from maximum extent.
CE9	57° 06' 4" N 3° 48' 6" W	0.53	Granite	Sub-angular, surface roughening. Some evidence of postglacial weathering	1.3	Crest of intermediate recessional Glen Einich lateral moraine	Thinning of Glen Einich ice from maximum extent.

Table 4.2: Glen Einich lateral moraine boulder descriptions

settlement hollows around the base of the boulder, a weathering band around the lowest parts of the boulder due to soil coverage, and splitting of the boulder due to frost activity. There is no modern shielding from trees, but the col between Glen Geusachan and Glen Lui was forested during the Holocene. The moraine stands below Carn a' Mhaim (1037 m), which provides a degree of horizon shielding.

4.1.3b: Intermediate moraine

Four samples were collected from a moraine ridge, which lies some 150 m to the west of the outer moraine (Figure 4.3 and table 4.4). It has similar profiles, dimensions and associations with channels. Between the two ridges are two other moraine ridge and channel systems, all continuous longitudinally for 400 to 600 m. GGb1, GGb2, GGb3 and GGb4 were taken from boulders of 2, 1, 1 and 2 m in height, sub-angular in shape and consistent with ice moulding on all surfaces. As with the boulders on the outer ridge, samples were chosen from boulders that showed no obvious signs of post depositional movement.

Figure 4.3: Glen Geusachan sampled boulders. Channel and ridge systems can be clearly seen on GGb2, GGb3 and GGb4, with GGb1 showing Glen Guesachan in the background.

Glen Geusachan marginal moraines

Sample	Location	Altitude	Lithology	Surface	Height	Context	Constraint Provided
GGa1	57° 07' 41"N 3° 46' 36" W	0.59 km	Granite	Sub-angular, surface roughening, evidence of postglacial weathering	1.2 m	Crest of outermost Glen Geusachan marginal moraine.	Extent of Glen Geusachan ice following ice sheet separation.
GGa2	57° 07' 49" N 3° 47' 5" W	0.59	Granite	Sub-angular, surface roughening, evidence of postglacial weathering	1.2	Crest of outermost Glen Geusachan marginal moraine.	Extent of Glen Geusachan ice following ice sheet separation.
GGa3	57° 01' 29" N 3° 47' 2" W	0.59	Granite	Sub-angular, surface roughening, evidence of postglacial weathering	1.8	Crest of outermost Glen Geusachan marginal moraine.	Extent of Glen Geusachan ice following ice sheet separation.
GGa4	57° 07' 56" N 3° 46' 42" W	0.59	Granite	Sub-angular, surface roughening, evidence of postglacial weathering	1.5	Crest of outermost Glen Geusachan marginal moraine.	Extent of Glen Geusachan ice following ice sheet separation.
GGb1	57° 01' 44"N 3° 46' 36" W	0.58 km	Granite	Sub-angular, surface roughening, evidence of postglacial weathering	2 m	Crest of intermediate Glen Geusachan marginal moraine.	Minimum extent of Glen Geusachan ice, prior to rapid retreat phase.
GGb2	57° 07' 44" N 3° 47' 5" W	0.58	Granite	Sub-angular, surface roughening, evidence of postglacial weathering	1	Crest of intermediate Glen Geusachan marginal moraine.	Minimum extent of Glen Geusachan ice, prior to rapid retreat phase.
GGb3	57° 07' 49" N 3° 47' 2" W	0.58	Granite	Sub-angular, surface roughening, evidence of postglacial weathering	1	Crest of intermediate Glen Geusachan marginal moraine.	Minimum extent of Glen Geusachan ice, prior to rapid retreat phase.
GGb4	57° 07' 56" N 3° 46' 42" W	0.58	Granite	Sub-angular, surface roughening, evidence of postglacial weathering	2	Crest of intermediate Glen Geusachan marginal moraine.	Minimum extent of Glen Geusachan ice, prior to rapid retreat phase.

Table 4.4: Glen Geusachan inner lateral moraine boulder descriptions

4.2: Ages

Cosmogenic age estimates for the Cairngorm boulder samples are displayed in Tables 4.5 and 4.6. The assumptions for the age estimates are outlined on pages 66 to 68.

4.2.1: The Glen More Moraine

Ages for the Glen More moraine vary between 13,177 ± 1418 a and 14,400 ± 1675 a, with a weighted mean age of 13,807 ± 701 a (Table 4.5). These ages are grouped into two distinct sets, the oldest being those on the moraine crest (CE3 and CE4). The samples from the moraine crest have a weighted mean age of 14,333 ± 1034 a, while the lower boulders have a weighted mean of 13,358 ± 954 a. The use of only two samples from each position is problematical, as only two results is statistically insignificant. However, the age difference make stratigraphical sense, in that the lower, younger boulders represent a thinner ice mass than the higher boulders on the crest. Moreover, the age measurements are internally consistent. Error margins from the two sets of samples here overlap. Thus the weighted means of the two sets of samples are within error, though the individual ages from each group are exclusive.

4.2.2: The Glen Einich lateral moraines

Ages for the Glen Einich lateral moraines vary between 12,513 ± 1296 a and 14,254 ± 1466 a, with a weighted mean age of 13,278 ± 675 a (Table 4.5). The outer moraine samples (CE8 and CE9) have a weighted mean age of 13,154 ± 924 a, and the inner moraine samples (CE6 and CE7) weighted mean age of 13,420 ± 987 a. The ages are within error and are internally consistent.

4.2.3: Glen Geusachan moraines

Ages from the Glen Geusachan lateral moraines vary from 11,711 ± 1285 a to 15,873 ± 1460 a, with a weighted mean age of 13,188 ± 529 a (Table 4.6). The two sampled moraines in this glen display ages that are not strictly in stratigraphic sequence. The samples from the outermost lateral moraine have a weighted mean age of 12,943 ± 739, and the inner moraine has a weighted mean of 13,447 ± 759. These weighted mean ages are within error and are not statistically separable.

Sample	Location	Moraine	Field Description	Sample Thickness	Altitude	S	T	P (at/g/yr)	10/9 Be Ratio	Error %	Atoms/g	Qtz Sample g	Age (ε = 0)	ε = 1mm/ka	ε = 3mm/ka	ε = 5mm/ka	ε = 1cm/ka	Weighted Mean (zero erosion)
CE1	57° 07' 49" N 3° 47' 2" W	Glen More	Proximal slope of Glen More moraine, on fluvioglacial ridge. 8 degree slope.	9 cm	0.43 km	0.996	1.501	7.50	0.229	5.6	102661	67.66	**13508**	13890	14254	14644	15760	
CE2	57° 07' 56" N 3° 46' 42" W	Glen More	Proximal slope of Glen More moraine, on fluvioglacial ridge. 8 degree slope.	2	0.48	0.996	1.607	8.19	0.182	7.5	107169	50.25	**13177**	12052	12324	12613	13423	Glen More Moraine
CE3	57° 07' 41"N 3° 46' 36" W	Glen More	Crest of Glen More moraine, on 4 degree northward slope. Open outlook	4	0.47	0.996	1.592	8.11	0.234	5.0	115165	58.77	**14293**	16594	17117	17687	19372	13807 ± 701
CE4	57° 07' 49" N 3° 47' 5" W	Glen More	Crest of Glen More moraine, on 4 degree northward slope. Open outlook	7	0.41	0.996	1.491	7.50	0.268	8.7	108635	68.97	**14400**	10961	11185	11422	12078	
CE6	57° 06' 23" N 3° 47' 59" W	Glen Einich	Crest of outermost lateral moraine, beneath Coire Creagach.	2	0.5	0.997	1.741	8.12	0.310	6.8	125817	73.91	**14254**	15163	15598	16068	17433	
CE7	57° 06' 12" N 3° 47' 59" W	Glen Einich	Crest of recessional lateral moraine, 100m east of outermost lateral.	3	0.5	0.997	1.741	8.08	0.262	7.1	112392	64.88	**12729**	13136	13460	13806	14790	Glen Einich Moraine
CE8	57° 06' 2" N 3° 48' 6" W	Glen Einich	Crest of recessional lateral moraine, 100m east of outermost lateral.	8	0.52	0.970	1.737	8.31	0.276	6.9	106699	78.78	**12513**	12702	13005	13328	14239	13278 ± 675
CE9	57° 06' 4" N 3° 48' 6" W	Glen Einich	Crest of outermost lateral moraine, beneath Coire Creagach. Open outlook	8	0.53	0.970	1.737	8.46	0.289	5.6	117791	75.09	**13817**	15975	16459	16984	18525	

S: Shielding Coefficient, measured in Radians of 15 degrees

T: Sample thickness correction. A depth dependent correction was employed as described by Masarik & Reedy (1995)

P: Production Rate, from Stone (2000) using scaling factors from Stone (1998), including atmospheric thickness and Muon Production variance corrections. Calculated at each sample site.

ε: Erosion rate (mm / ka).

Table 4.5: Summary of cosmogenic data from Glen Einich.

Sample	Location	Moraine	Field Description	Sample Thickness	Altitude	S	T	P (at/g/yr)	10/9 Be Ratio	Error %	Atoms/g	Qtz Sample g	Age (ε = 0)	ε = 1mm/ka	ε = 3mm/ka	ε = 5mm/ka	ε = 1cm/ka	Weighted Mean (zero erosion)
GGa1	57° 07' 41"N 3° 46' 36" W	Geusachan Outer	Crest of outermost Glen Geusachan lateral moraine.	8 cm	0.57 km	0.998	1.686	9.26	0.194	7.8	100228	39.79	**11711**	10952	11176	11412	12067	
GGa2	57° 07' 49" N 3° 47' 5" W	Geusachan Outer	Crest of outermost Glen Geusachan lateral moraine.	10	0.57 km	0.998	1.657	6.47	0.295	4.1	115810	58.1596	**13780**	18283	18922	19624	21745	12942 ± 739
GGa4	57° 07' 56" N 3° 46' 42" W	Geusachan Outer	Crest of outermost Glen Geusachan lateral moraine.	7	0.57 km	0.998	1.701	8.44	0.226	6.8	114445	43.04	**13258**	13771	14128	14511	15606	
GGb1	57° 01' 44"N 3° 46' 36" W	Geusachan Inner	Crest of innermost Glen Geusachan lateral moraine.	7	0.56	0.998	1.686	7.07	0.253	5.0	135731	42.62	**15873**	19624	20363	21182	23699	
GGb2	57° 07' 44" N 3° 47' 5" W	Geusachan Inner	Crest of innermost Glen Geusachan lateral moraine.	7	0.56	0.998	1.686	8.93	0.217	7.4	116423	41.71	**13608**	13232	13561	13913	14912	13447 ± 759
GGb3	57° 07' 49" N 3° 47' 2" W	Geusachan Inner	Crest of innermost Glen Geusachan lateral moraine.	9	0.56	0.998	1.657	7.36	0.317	5.4	100190	75.58	**11917**	13828	14188	14574	15679	

S: Shielding Coefficient, measured in Radians of 15 degrees
T: Sample thickness correction. A depth dependent correction was employed as described by Masarik & Reedy (1995)
P: Production Rate, from Stone (2000) using scaling factors from Stone (1998), including atmospheric thickness and Muon Production variance corrections. Calculated at each sample site.
ε: Erosion rate (mm / ka).

Table 4.6: Summary of cosmogenic data from Glen Geusachan.

4.3: Establishing Criteria for Age Estimates

4.3.1: Erosion

All ages discussed have been stated in terms of 'zero erosion', however this is clearly not an accurate representation of reality. It is widely accepted that exposed rock surfaces weather due to the action of the elements, granite in the Scottish environment being particularly susceptible to frost action, and chemical dissolution (hydrolysis) of feldspars as a result of rainwater. Quartz also may be dissolved by the addition of water ($SiO_2 + H_2O = Si(OH)_4^0$), but this is an extremely slow process and not significant at the timescales under consideration in this study. The hydrolysis of feldspars in comparison, is relatively rapid, leading to the structural breakdown of the granite, and the production of gibbsite amongst other products, depending on the original composition of the granite. As a result quartz grains at the surface of the boulder are frequently detached. This product forms a gravely soil, or 'gruss' surrounding the boulder. Indeed much of the plateaux surfaces of the Cairngorms are covered in gruss material, indicating a potentially high erosion rate.

Few studies have been completed that have attempted to quantify the rate of rock weathering, and produce an estimate of rock surface removal, and even fewer of these have concentrated on the effects of weathering on granite surfaces (Hampel *et al.* 1975, Nishiizumi *et al.* 1986, 1991, Bierman, 1994). Stroeven *et al.* (2002) argue for rates of around 1.6 mm / ka on tor forms in central Scandinavia, however these rates are not representative of interglacial or interstadial erosion rates, as the authors argue for the preservation of tors beneath cold based ice during glacial phases. Cosmogenic ^{10}Be and ^{26}Al age estimates from Cairngorm tor samples show that both the tors and the weathering pits they display on their upper surfaces, have complicated exposure histories, dating back at least 50 ka. This and the elevation of quartz veins above granite surfaces, has led researchers to conclude that weathering rates on Cairngorm summits are in the region of 10 mm / ka (Hall, *pers comm*).

If one takes the 10 mm / ka erosion rate for the plateau summits as a maximum possible rate, and uses the 'zero erosion' rate as a baseline minimum, then a range of possible rates, and concomitant ages may be produced (Tables 4.5 and 4.6). Hall (*pers comm*) argues that erosion rates in Cairngorm glens should be significantly lower that those on the summits. Therefore likely erosion rates are those in the 5 to 10 mm / ka range. An erosion rate of 5 mm / ka produces mean ages of Cairngorm moraines of around 15 to 13.6 ka BP, whilst the higher end of the scale, a rate of 10 mm / ka produces mean ages of 16.6 to 15 ka BP for the various moraine groups (Table 4.7).

Moraine	Zero Erosion (EWM age)	5 mm / ka (EWM age)	10 mm / ka (EWM age)
Glen More Crest	13,928 ± 1029	14,333 ± 1034	16,584 ± 1197
Glen More Retreat	13,358 ± 954	13,599 ± 973	15,284 ± 1091
Glen Einich Laterals	13,278 ± 675	14,828 ± 756	15,171 ± 771
Glen Geusachan	13,188 ± 529	14,991 ± 612	15,035 ± 605

Table 4.7: Moraine groups error weighted mean ages.

4.4: Possible other sources of variance in age estimation

4.4.1: Use of Different Production Rates

As has been discussed, published production rates vary greatly from 4.74 atoms g^{-1} yr^{-1} (Clark *et al.* 1995) to 6.4 atoms g^{-1} yr^{-1} (Brown *et al.* 1991). The use of different rates to that of Stone (1998) will inevitably produce age estimates for these samples that may be anything up to ±1 ka different from the ages described here.

4.5: ^{14}C Dating of Loch Etteridge Sample

The core taken from Loch Etteridge revealed a complex basal signal (Figures 4.3 and 4.4). The basal 50 cm was characterised by fine inorganic clays, conformably overlying fluvial gravels. Within the clay sections were three layers of clay-gyttja, the first and earliest stratigraphically, at 5 cm above the gravel and 8 cm thick, the second at 14 cm above the gravel and 7 cm thick, and the youngest stratigraphically at 28 cm above the gravels and 5 cm thick. The uppermost clay section was 14 cm thick and conformably overlain by a 7 m section of peat deposits. Within the youngest gyttja section and the topmost clay layer two distinct tephra deposits were found.

4.5.1: AMS Radiocarbon Results

A basal AMS radiocarbon age of 12,930 ± 40 a (15,246 ± 713 calendar a BP, calculated using Calib 4.3) was produced for Loch Etteridge (Figure 4.3 and 4.4). This is in good agreement with the age determined by Sissons and Walker (1974), and combined with the stratigraphical consistency of the other ages determined by the 1975 study it provides a minimum age for the deglaciation of this part of the Spey Valley (Table 4.9).

4.5.2: Tephrostratigraphical Results

Two microscopic tephras were found in the basal section of the Loch Etteridge core, the lowest (and therefore earliest stratigraphically) at 734 cm, and the second at 714-717 cm (Figure 4.4). Their position in the core, relative to sections of minerogenic material, and their morphologies, argued that these may represent the Borrobol Tephra (12.5 ^{14}C ka) and the Vedde Ash (10.6 ^{14}C ka) (Roberts, *pers comm*, Roberts *et al.* 1997, 1998, Lowe & Turney *et al.* 1997, Turney *et al.* 1998) (Figure 4.3). The basal ^{14}C age from this core similarly provided an earliest constraint to tephra deposition. The paucity of shards within the core prevented a full geochemical analysis from being carried out to determine the exact provenance of the tephras.

If these two signals do represent the Vedde and Borrobol tephras, then their stratigraphical position and ages support both the new and the 1975 radiocarbon dates. Calibration of the dates of these tephras give ages of 12.4 – 12.9 ka BP for the Vedde Ash, and 15.5 - 14.1 ka BP for the Borrobol tephra. The basal calibrated radiocarbon date is older than both of these ages, providing further constraint to the cosmogenic ages from the Cairngorms.

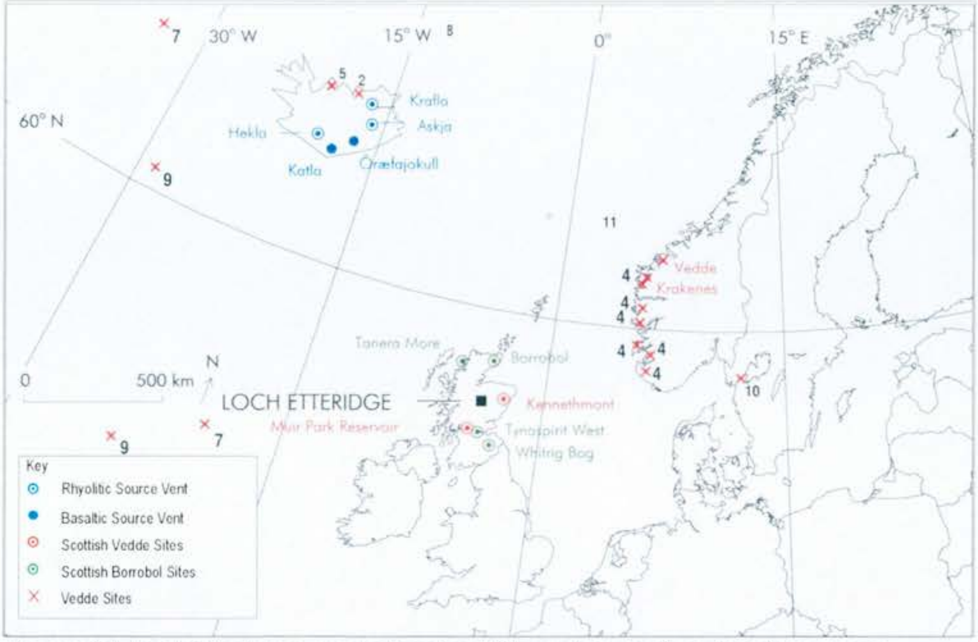
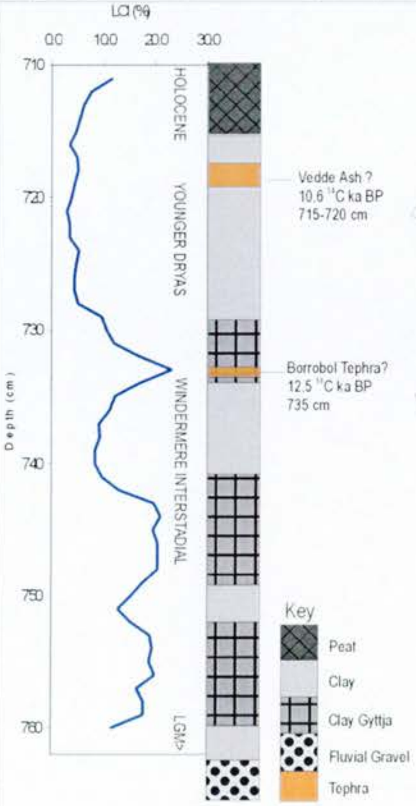
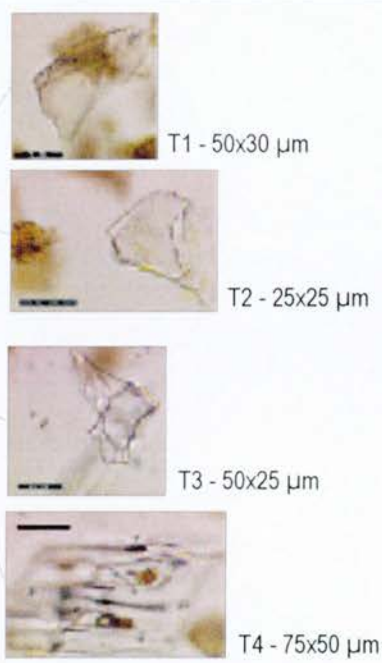


Figure 4.3: Loch Etteridge, and other Scottish and Scandinavian sites where tephras have been found, with probable Icelandic sources.





Sample	Depth	Date
SRR-301 (1975)	6.97 – 7.01 m	9,405 ± 260
SRR-302 (1975)	7.14 – 7.15 m	10,764 ±120
SRR-303 (1975)	7.25 – 7.26 m	11,290 ± 165
SRR-304 (1975)	7.29 – 7.32 m	13,151 ± 390
LED1 (2000)		12,930 ± 40

Table 4.9. Summary of radiocarbon dates from Walker (1975), and Everest & Roberts (in prep).

Figure 4.4: Loch Etteridge Core D1, with associated Tephras, Loss on Ignition % values, and possible Quaternary periods.

Chapter : 5

Interpretation and Discussion

Chapter 5: Interpretation and Discussion

5.1: Introduction

This chapter discusses the results of the cosmogenic ages in terms of the interactions of the Cairngorm Ice Cap and the Scottish Ice Sheet, though issues arising from the cosmogenic technique must be clarified before any discussion of absolute ages can be made. The nature of the cosmogenic technique depends on the presentation of results in terms of erosion rate. As a result this chapter focuses on the means by which ages can be best presented. Two sets of conclusions will be reached; the first highlights interpretations of the results that can be made independent of erosion rate corrections; the second set of conclusions focuses on absolute ages that can be constrained by data collected from other sources.

Transparency of the results is best achieved through their presentation in terms of a range of ages, from the youngest possible, i.e. those with a zero rate of erosion applied to them, to the oldest, i.e. those which have been derived using an upper limit of erosion calculated by independent constraints. This provides an envelope of possible ages.

The ages determined for the three sets of moraines in this study have been assigned an envelope of erosion rates from zero to a maximum of 10 mm / ka. The age of each moraine has been calculated using the error weighted mean of the ages of all boulders upon it. The ages are presented in Table 5.1.

Erosion	Glen More	Glen Einich	Glen Geusachan
0	13,806 ± 701	13,278 ± 675	13,188 ± 529
10 mm / ka	15,874 ± 806	15,171 ± 771	15,035 ± 605

Table 5.1: Zero erosion and 10 mm / ka erosion, error-weighted mean ages for the three moraine groups.

5.2: Conclusions independent of erosion rate uncertainties

All ages stated in this section are based on zero erosion of the samples. As a result they are absolute minimum ages. Isotopically it is impossible for the ages to be younger than presented here.

5.2.1: Correlation of the Glen More and Glen Einich Moraine Ages

The cosmogenic data from both sets of moraines in Glen Einich argue for synchrony between the local and regional ice masses. Baseline minimum ages derived from zero erosion calculations reveal that the Glen Einich glacier and Glen More lobe were contemporaneous between 13.8 and 13.3 ka BP. Further analysis of the different areas of the Glen More moraine show different ages for the crest, and for a lower retreat position on its northern flanks (Table 4.7). The crest of the Glen More moraine has an age of 14.3 ± 1 ka, at a time when ice was presumably still extensive in Glen Einich. The weighted mean age of the retreat position on the flanks of the Glen More moraine (13.4 ± 1 ka BP) shows that thinning of the Glen More lobe was contemporaneous with the age of the Glen Einich moraines (13.3 ± 0.7 ka BP). The key conclusion here is that ice in Glen More occupied the moraine position at the mouth of Glen Einich at the same time that ice in Glen Einich occupied the dated moraine positions beneath Coire Creagach. In other words the Scottish Ice Sheet margin in Glen More and the Cairngorm Ice Cap limits in Glen Einich are synchronous.

5.2.2: Correlation of Moraines in Glen Geusachan, Glen Einich and Glen More

Zero erosion weighted mean ages of the Glen Geusachan moraines of 13.2 ± 0.5 ka BP, are almost identical to the ages for the Glen Einich final retreat position, at 13.3 ± 0.7 ka BP. This argues for a synchronous stage of ice in both glens. Using the same logic the outer moraines in Glen Geusachan are the same age as the Glen More moraine at the mouth of Glen Einich. The significant conclusion is that the stillstand identified in the geomorphology of the Cairngorms, and illustrated by the moraines in Glen Einich, Glen Geusachan and Glen More, occurred at the same stage during deglaciation, and was common to both local and regional ice masses.

5.2.3: Stillstand of ~1 ka of local and regional ice masses.

Zero erosion error weighted means from the crest and retreat positions on the Glen More moraine of 14.3 ± 1 and 13.5 ± 1, imply that the Spey ice lobe in the Glen More basin could have remained at this marginal position for around 1 ka. Interestingly, this conclusion is supported by two independent lines of evidence. First, if the age range from the highest / outer moraines and lower / inner moraines are accepted, then the glacier surface fell from around 550 to 450 m in 800 years. The relationship of the lake features to the various ice limits demonstrates the persistence of a lake throughout the period of thinning, implying overall glacier stability. Secondly, the couplets within the Glen Einich sedimentary section reveal a sequence of some 400 - 800 annual clay / silt pairs (Brazier, *et al.* 1998) (Figures 3.13, 3.14 and 3.15). These also are consistent with the persistence of a lake.

The main conclusion that can be drawn from the interpretation of the Glen Einich geomorphology and lake sediments is that the Glen Einich lake, and thus the Glen More lobe occupied a stable position at the mouth of Glen Einich for a period of about 1 ka.

5.2.4: Reconstruction of a stage in Cairngorm Ice Cap Deglaciation

Major valleys such as Glen Luibeg, Glen Derry, Glen Dee and Glen Einich, as well as the Glen More basin, display large moraines. It is argued that a staged retreat pattern can be identified in the massif, with the final phase providing the dating evidence shown here. Figure 5.1 shows a reconstruction of glacial limits of the Scottish Ice Sheet ice in Glen More at several stages during its decline. The figure also shows the final phase of stability of both the Glen More ice and the Cairngorm Ice Cap.

Geomorphology and ice-dammed lake evidence from the Lairig Ghru and the Glen More basin, indicate that there were several retreat stages. The lacustrine evidence from the Lairig Ghru indicates a lake dammed by ice in Glen More at a higher stand than that in Glen Einich (Figure 5.1). Furthermore at such a stage meltwater from Glen Einich drained over the col south of Carn Eilrig, into the Lairig Ghru lake. Though no dating has been carried out here, the geomorphology indicates that this situation persisted for a considerable period of time, as the lake sediments in the Lairig Ghru are extensive (Brazier *et al.* 1998). Further east along the northern flanks of the massif in the Glen More basin, large meltwater channel features and kame terraces at higher elevations reveal even earlier phases of stability during the thinning of the Glen More lobe.

The final phase of retreat is characterised by the decoupling of Cairngorm ice from that of the main Scottish Ice Sheet. The Glen More lobe had thinned significantly by this point, and its marginal position is marked by lateral channels sweeping round the lower slopes of the Cairngorms. The Cairngorm Ice Cap maintained limits close to those shown in Figure 5.1 throughout this period. Valley glaciation was confined solely to Garbh Coire, and to Glens Einich, Geusachan and Eidart. Moraines in all these locations support this occupation. There are no other significant troughs in the western Cairngorms draining the western plateau. The plateau itself was ice covered, supplying the valley glaciation, plus limited glaciation of areas such as the col to the north of Braeriach, and the corries above the lake dammed between the Glen Geusachan and Garbh Coire glaciers. Analogous situations to this phase persist in present day Greenland, the Canadian high Arctic and arctic Norway where large numbers of small plateau ice caps supply ice to a few valley and corrie glaciers (Manley, 1955, Rosqvist & Ostrem, 1989, Rea *et al.* 1998, 1999) (Figure 5.2).

5.3: Absolute Ages of the Cairngorm Stillstand

5.3.1: The Loch Etteridge ^{14}C Constraint

Two basal dates have been acquired from the Loch Etteridge site (15.2 ± 0.7 and 15.5 ± 1.1 cal ka BP) which are in close agreement with each other, despite the use of different measurement methods. It is envisaged that these data may be used as an independent age constraint for the Glen More moraine at the mouth of Glen Einich. The site is the nearest to the Cairngorms that has yielded a core suitable for ^{14}C dating, despite repeated attempts to find one closer. Coring for organic sediments predating the Holocene has been undertaken in Glen More without success, and so it was decided to re-test the date derived from an earlier study (Sissons and Walker, 1974, Walker, 1975). Loch Etteridge occupies a position in Glen Truim, a tributary of the Upper Spey, and ^{14}C ages from the site provide minimum dates for the wastage of the Late Devensian ice sheet from the surrounding area (Walker, in Gordon and Sutherland, 1993). In particular the ^{14}C ages from this site provide the only constraint for ice retreating south west up the Spey from Glen More.

As Loch Etteridge lies 'up-ice' in the Spey Valley, any age derived from it should be younger than glacier margin positions in the Glen More basin. Indeed if the ^{14}C ages are to be accepted, then the moraine ages in the Cairngorms cannot be younger than around 15.2 – 15.5 ka BP, as organic sediment accumulation at Loch Etteridge can only have begun once the ice had disappeared from the site. Abandonment of the Glen More moraine at the mouth of Glen Einich must have taken place earlier than this date.

There are several qualifications to consider when considering the Loch Etteridge dates. The geomorphology surrounding Loch Etteridge is symptomatic of a site formed at or beneath the margin of a downwasting ice mass (Figure 5.3). The esker and kame terrace morphology reflect formation as a result of stagnant ice (Bennett and Glasser, 1996), and it is conceivable that the Etteridge core site was not covered by glacial ice at a time when ice was still present in the Truim and Tummel valleys. The presence of buried ice in the environment may also have led to the production of a falsely young date from the site. Some significant time is required for the buried ice masses to melt out before organic matter accumulation may begin (Everest and Bradwell, *in press*).

The distance between Loch Etteridge and the margins of the Glen More lobe is around 25 km. Assumptions have to be made about the time taken for retreat to take place up the Spey Valley from Glen More to Loch Etteridge, and again this implies an interval between the abandonment of the Glen More moraines and the deglaciation of Loch Etteridge.

A major boost to the interpretation of the ^{14}C dates from Loch Etteridge is the discovery of two microscopic tephras above the basal section of the core. These are believed to represent the Borrobol Tephra and Vedde Ash (15.5 - 14.2 and 12.9 - 12.6 ka BP respectively). Their positions in the core and in combination with the basal dates argue for stratigraphic integrity and they also provide an independent constraint to the ^{14}C data.

Therefore the Loch Etteridge dates are independent dated constraints on the deglaciation of the Cairngorms. It must be argued that accumulation of organic sediments at Loch Etteridge postdated the formation of the Glen More moraine. Thus the age of the last Cairngorm stillstand involving the Glen More ice and ice in Glen Einich and Glen Geusachan is at least 15.2 ± 0.7 ka BP.

5.3.3: Erosion Rate Corrections

By employing the Loch Etteridge ^{14}C constraint to the Cairngorm sample weighted mean ages, assumptions concerning predicted erosion rates may be made. The application of an erosion rate of 10 mm / ka to all ages provides the best fit to the radiocarbon minimum age constraint. In general it has the effect of offsetting the cosmogenic ages by around 2 ka in most cases. Sensitivity testing reveals that age increases linearly with erosion, until around 25 mm / ka for all samples. Upwards of this value the erosion rate reaches steady state, and ages tend to infinity (Figure 5.4).

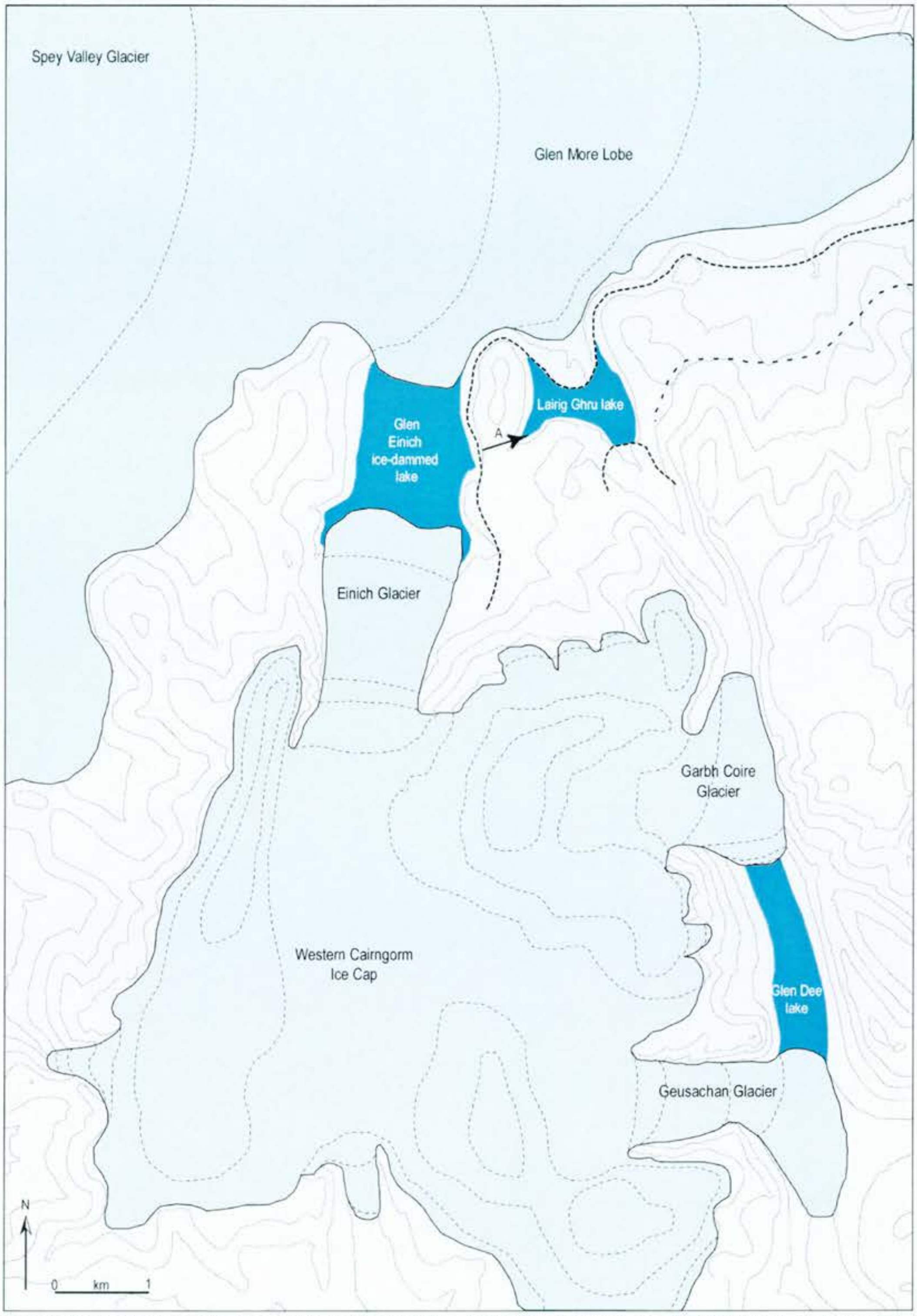


Figure 5.1: A diagrammatic reconstruction of the Western Cairngorm Ice Cap at 15.5 ka BP. Limits dated by cosmogenic ^{10}Be can be correlated between Glen Einich, Glen More and Glen Geusachan. Earlier limits (thick dotted lines) can be seen outside the main Glen More lobe limits shown here, indicating staged thinning of the Glen More lobe, concurrent with retreat of the Cairngorm ice sheet systems, seen in glens such as Lui and Derry.

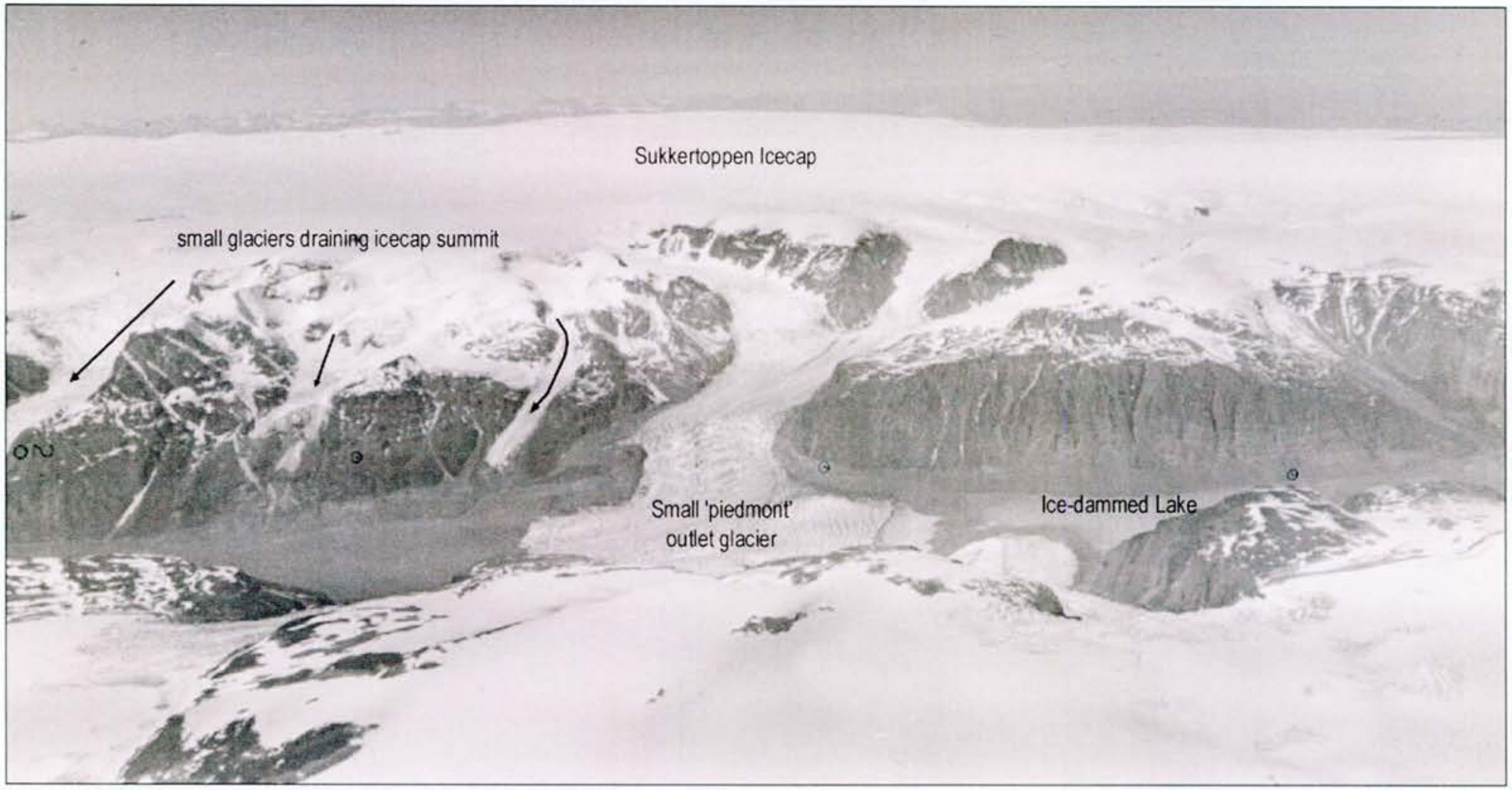


Figure 5.2: Sukkertoppen Icecap in West Greenland. An illustration of an analogous situation to Glen Geusachan during the deglacial phase. A small icecap, situated on a flat plateau summit supplies ice to the largest valleys draining it. Smaller glaciers sit around the break of slope from the plateau into the adjoining valley. The 'piedmont' glacieris one of only a small number of large glaciers draining this plateau, and is sufficiently large enough to have dammed a lake within the main valley (photo Dansk Geodetisk Institut).

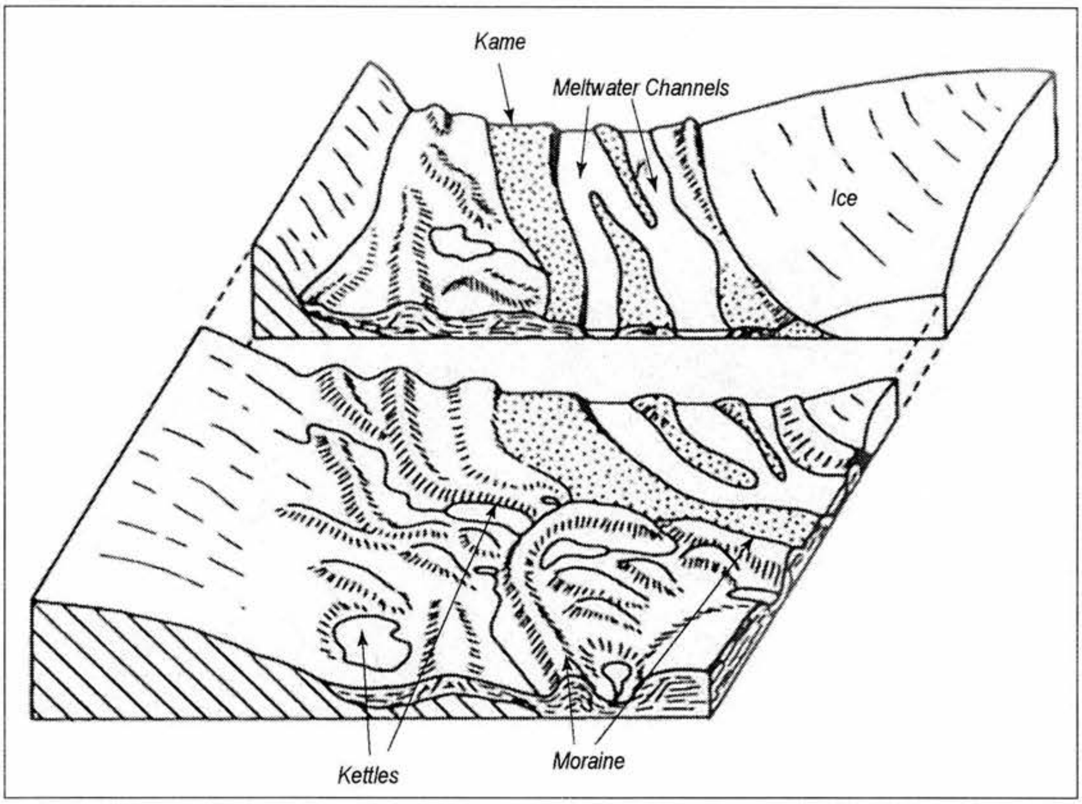


Figure 5.3: Schematic representation of an environment of glacial retreat, dominated by fluvioglacial activity (after Bennett and Glasser, 1996)

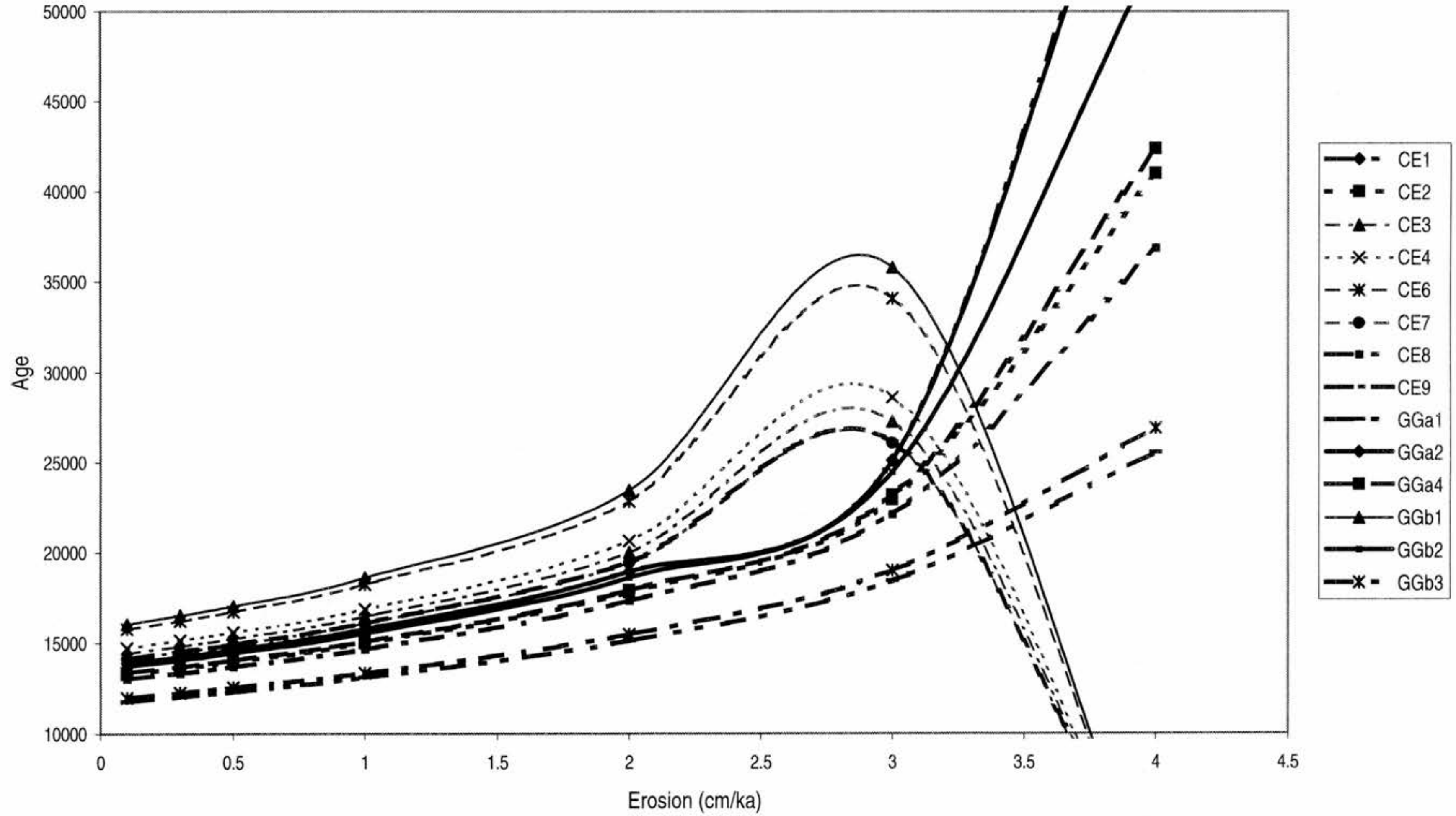


Figure 5.4: Sensitivity test results for the Cairngorm ^{10}Be samples. The model predicts that saturation will occur at around 2.5 - 2.7 cm of erosion per ka. Beyond this limit ages steady state erosion is reached and ages trend to infinity (dips on curves are due to software).

By employing a 10 mm / ka erosion rate the weighted mean ages of the Glen More, Glen Einich and Glen Geusachan moraine groups are 15.9 ± 0.8, 15.2 ± 0.8 and 15 ± 0.6 ka BP. The envelope of ages thus produced for the Cairngorm moraines is therefore summarised in Table 5.2. If the ^{14}C ages from Loch Etteridge are accepted, then the 10 mm / ka erosion rate appears to be the most likely, indeed erosion rates for Cairngorm summits have been proposed to be of the order of 10 mm/ ka (Hall, *pers comm.*).

Moraine	**0 mm / ka (EWM Age)**	**10 mm / ka (EWM age)**
Glen More Crest	14,333 ± 1034	16,584 ± 1197
Glen More Retreat	13,358 ± 954	15,284 ± 1091
Glen Einich Laterals	13,278 ± 675	15,171 ± 771
Glen Geusachan	13,188 ± 529	15,035 ± 605

Table 5.2: Erosion corrected error weighted mean ages of major moraine groups in the Cairngorms. These will form the basis for the correlations of deglaciation of the region in the following sections.

Chapter 6:

The Wider Picture- Deglaciation of the Cairngorms in a Scottish Context

Chapter 6: The Wider Picture- Deglaciation of the Cairngorms in a Scottish Context

6.1: Lateglacial Fix on the eastern Scottish Ice Sheet and Cairngorm ice

For the first time an age has been established for a stage of deglaciation of the main Scottish Ice Sheet and the smaller local Cairngorm ice cap. For a period of around 1 ka the overall retreat of Scottish ice ceased and there was a period of stability.

During the 1 ka stillstand the evidence in Glen Einich shows the ice thinned by around 50 to 100m. Prior to this the ice edge retreated by 50 km from the North Sea coast, and 120 km from limits offshore. Subsequently it retreated to expose the whole of Scotland within another 1 ka. Immediately prior to the stillstand event, the Glen More lobe fell from positions high on the northern flanks of the Cairngorms at the Chalamain Gap, and Airgiod Meall down to the site of the Glen More moraine, with at least two stillstands preceding the one dated here. These three stillstand events are marked by moraine, meltwater channel and ice dammed lake features. The earliest event is shown by the moraine and channel features running along the northern flanks from the Airgiod Meall, which are continuous into the Pass of Ryvoan. The second event caused ice in Glen More to dam a lake in the Lairig Ghru, shown by extensive lacustrine sediments. The final event dammed a lake in Glen Einich for a period of around 1 ka. For at least the final event evidence suggests that local ice in the Cairngorms also halted its decline at the time, in a similar fashion to the Scottish Ice Sheet in Glen More.

6.2: Correlation with North Atlantic Climate changes- the role of Heinrich Event 1

From examining all the relevant data collected in this study it is possible to begin to correlate events at the local scale of the Cairngorms with the wider North Atlantic. The deglacial geomorphology suggests that the stillstand was the final cold period immediately prior to rapid retreat up the Spey Valley, and perhaps prior to the final deglaciation of Scotland. The cosmogenic data, constrained by erosion rate and independent ^{14}C dates, place an age on this period of between 15 and 16 ka BP.

The evidence places the stillstand seen here at the end of the cooling period immediately preceding lateglacial warming, which saw temperatures attain present day values. This pattern is well displayed in the GRIP and GISP2 ice cores (Figure 6.1). The spread of 10 mm/ka erosion-corrected cosmogenic ages broadly correlates with

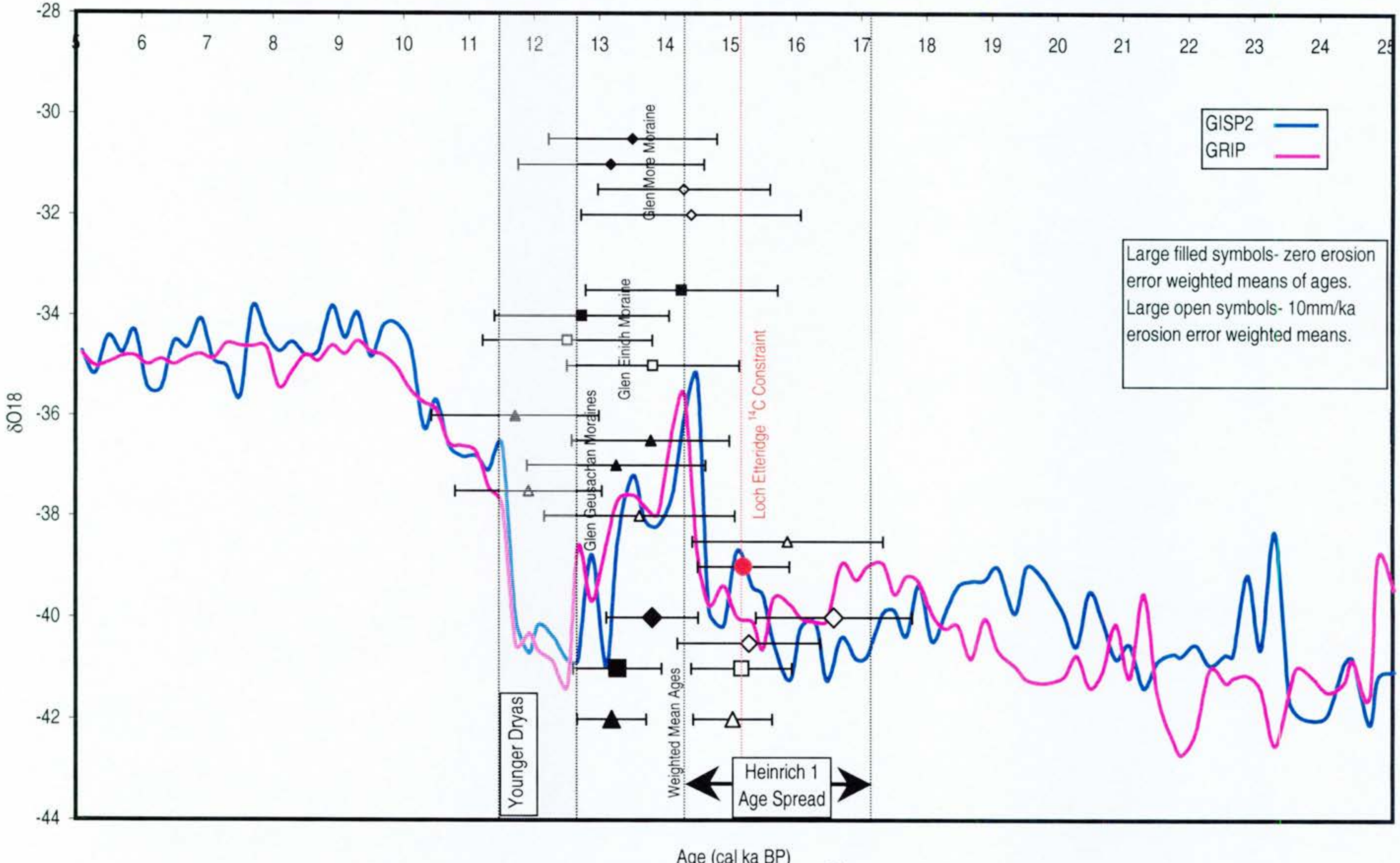


Figure 6.1: Greenland Ice Core records and Summary of Cairngorm cosmogenic ^{10}Be ages, with error weighted mean ages plotted as larger symbols. Loch Etteridge ^{14}C constraint shown as red symbol.

the Heinrich-1 cold event dated to 18.1 to 15.9 ka BP (Bond *et al.* 1992, 1993, Ballantyne *et al.* 1998, Keigwin & Lehman, 1994, Broecker, 1994, Andrews, 1998, Maslin *et al.* 1995, McCabe *et al.* 1998, 2002). In fact even the zero erosion error weighted means of all samples correlates with the youngest published ages for H-1, or the sudden cooling periods seen in the GISP and GRIP records at around 13 and 14 ka BP. The geomorphology of the Cairngorms therefore reflects an ice sheet response that coincides with a climatically driven environmental change. Sudden discharges of ice into the North Atlantic lowered temperatures in the region during Heinrich events and saw greater extents of sea ice. The temperature drop was sufficient to replenish glaciers around the margins of the North Atlantic and halt retreat. The damming of lakes represented in Figure 5.1 is the Cairngorm equivalent of such a stage. The rapid retreat subsequently would have taken place during the dramatic warming that followed Heinrich-1.

The range of ages produced by erosion rate calculations and the uncertainty in dating Heinrich-1 makes any tighter correlation impossible. However the geomorphological evidence agrees in that it points to a pattern of events identical in trend and amplitude to that identified in the ice cores. The fact that separate ice masses of different size, the Scottish Ice Sheet represented by the Glen More lobe and the local Cairngorm Ice Cap, responded similarly, implies that there was a change in the direct climatic signal that affected all glaciers, such as would be achieved by a temperature fall.

Further strengthening of this correlation between the H-1 event, and the Cairngorm stillstand is provided by the ^{14}C data from Loch Etteridge. Despite the limitations of the site in terms of its proximity to the Cairngorms and its position within the Truim valley, it still provides the best dated constraint to the abandonment of the Spey Valley by the Scottish Ice Sheet. The latest possible time for ice to have occupied the Glen More basin is immediately prior to 15.6 ka BP. If one accepts that deglaciation was rapid after this time, then an H-1 cause for the stillstand in the Cairngorms and the Spey Valley is likely. Indeed the stepped change in the rate of deglaciation implied by the geomorphology argues for such a correlation.

The earlier halts in retreat of the Scottish Ice Sheet identified on the northern flanks of the Cairngorms in Glen More and by the lake evidence in the Lairig Ghru represent earlier cooling events during the overall deglaciation. The ice core records display two cooler periods prior to the final cold snap attributed to H-1, which occur around 1 and 2 ka earlier. It is tempting to relate the earlier stillstands to the cooler periods at 16 and 16.6 ka BP.

6.2.1: ELA Changes of the Cairngorm Ice Cap

Reconstruction of ELAs on the Western Cairngorm Ice Cap constrains the cooling required to nourish the stillstand (Figure 6.2 and 6.3a, b and c). The AAR method was chosen to reconstruct the ELA in accordance with the suggestions of Torsnes *et al.* (1993), employing an AAR of 0.6 ± 0.05. Ice divides were used as the boundaries between areas of accumulation, these being determined by ice surface gradient. Reconstructing palaeo ELAs presents significant problems when comparing glaciers that have different hypsometries, particularly in the snout areas. For example a glacier with a wide accumulation area and a narrow snout may have a different AAR than one with a small accumulation area and wide snout, even though their ELAs may be the same (Furbish & Andrews, 1984). Therefore employing a uniform AAR value for a number of glaciers with different hypsometries may produce results with significant errors (Benn & Evans, 1998), and consequently secondary measures must be employed as a check. The Balance Ratio Method, described by Furbish and Andrews (1984) takes account of both glacier hypsometry, and the shape of the mass balance curve, however some knowledge of the rates of accumulation and ablation must be assumed, and so therefore this measure has not been used in this study. A far more simple measure of the altitude of the upper limit of ablation is the Mean Elevation of Lateral Moraines (MELM) method (Torsnes, *et al.* 1993). Reconstructed ELA and MELM values for the three largest outlet glaciers of the Western Cairngorm Icecap are summarised in Table 5.1.

Glacier	Total Area (km^2)	ELA (m)	MELM (m)
Glen Einich	14	865 ± 43	650
Glen Geusachan	15.75	872 ± 44	630
Garbh Coire	6.95	921 ± 46	940

Table 6.1: Summary of ELA values for main valley glaciers draining the Western Cairngorm Ice Cap

The three valley glaciers exhibit very similar ELA values of 865, 872 and 921 m for Glen Einich, Glen Geusachan and Garbh Coire respectively (errors are derived from the AAR method) and MELM values of 650, 630 and 940 m. The seemingly low MELM values in Glen Einich and Glen Geusachan are thought to be the result of the steepness of the valley sides in the upper regions of the palaeo ablation areas, resulting in little or no lateral moraines being preserved due to Holocene slope mass movement processes. One significant factor that cannot easily be taken into account in the reconstruction of ELAs is past precipitation levels. Most commonly ELA studies present results in terms of temperature depression from present day levels. Present day annual mean temperature data for the Cairngorm summits varies between 1 and 0°C, placing the modern day ELA of the western Cairngorms at around 1300 - 1400 m. This would imply that a 2 - 3°C temperature depression from today's values existed during this phase of deglaciation, assuming an adiabatic lapse rate of 0.6° per 100 m altitude.

6.3: The Loch Lomond Stadial in the Cairngorms

By determining consistent ages for each of the three moraines in this study, new interpretations can be made concerning the impact of the Loch Lomond Stadial on the Cairngorms. The long-discussed valley glaciation of Glen Geusachan, Garbhe Coire and Glen Eidart (Sissons, 1979, Sugden, 1980, Bennett and Glasser, 1991) now appears to belong to an earlier period during the main deglaciation of the massif, rather than to glaciation during the Younger Dryas. As the limits correlate with each other across the massif, both in terms of zero erosion ages, and erosion corrected ages, there can be little doubt that the glaciers in Glen Einich and Glen Geusachan existed at the same time as ice from the main Scottish Ice Sheet filled the Glen More basin.

The difficulty that authors have experienced in placing the deglaciation of Glen Geusachan into a Loch Lomond Stadial chronology, in the light of the rapid climatic changes of the time seems well founded (Bennett and Glasser, 1991). The geomorphology of the glen fits with the popular British model of Loch Lomond Stadial glacier decline in that rapid retreat theoretically leads to the construction of 'hummocky moraine' (Sissons, 1979, Bennett, 1991, Bennett and Boulton, 1993a, 1993b). However, there is little other evidence of 'hummocky moraine' in the rest of the Cairngorms, save that at Loch Builg, which has been attributed to a period of deglaciation during the Late Devensian (Clapperton *et al.* 1975). The cosmogenic ages from the lateral moraines in Glen Geusachan argue for the disappearance of ice from the glen prior to the commencement of the Younger Dryas. Even though two of the 'zero-erosion' ages (GGa1 and GGb3) fall within the period associated with Younger Dryas glaciation, these have been interpreted as outliers to the other

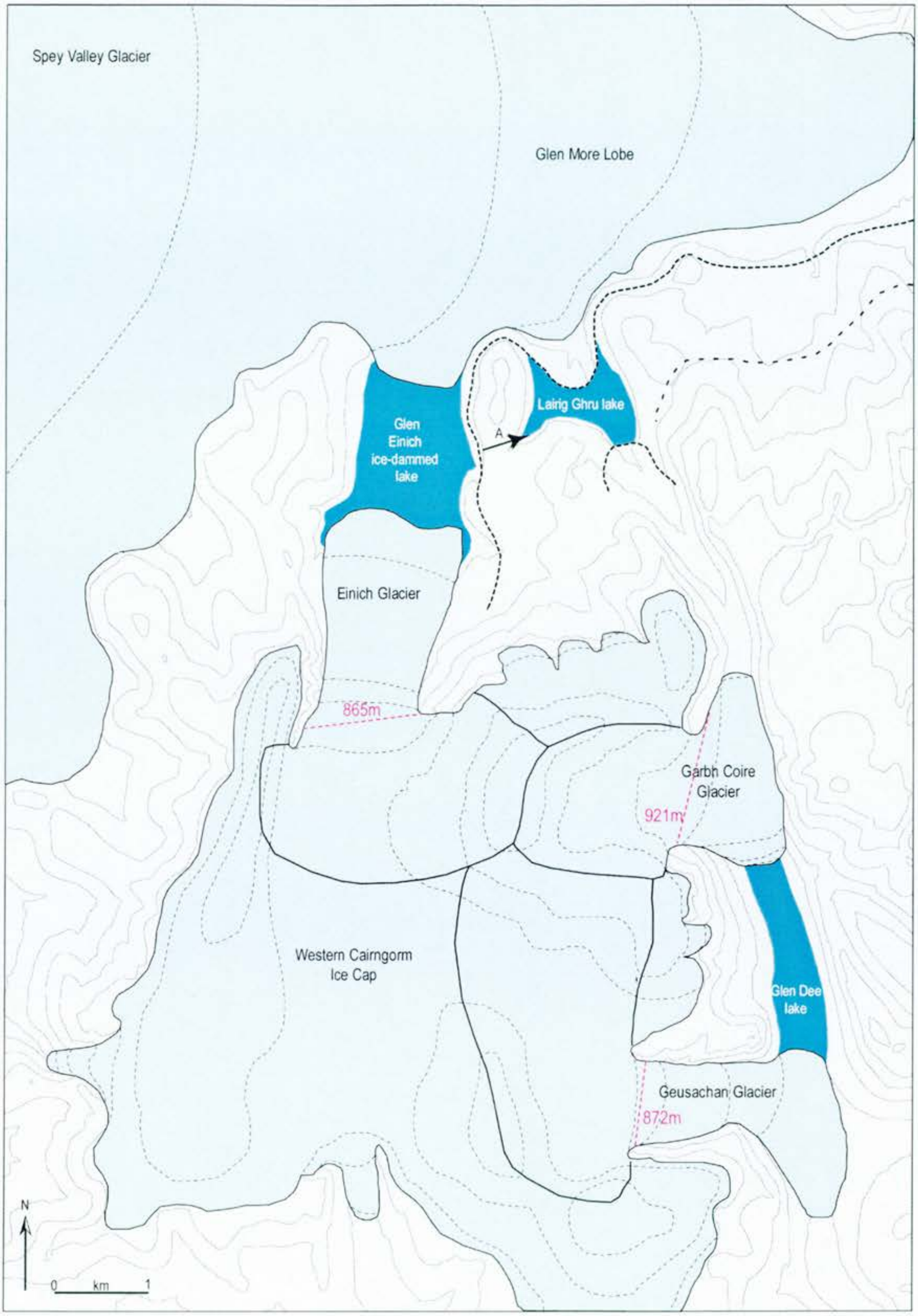
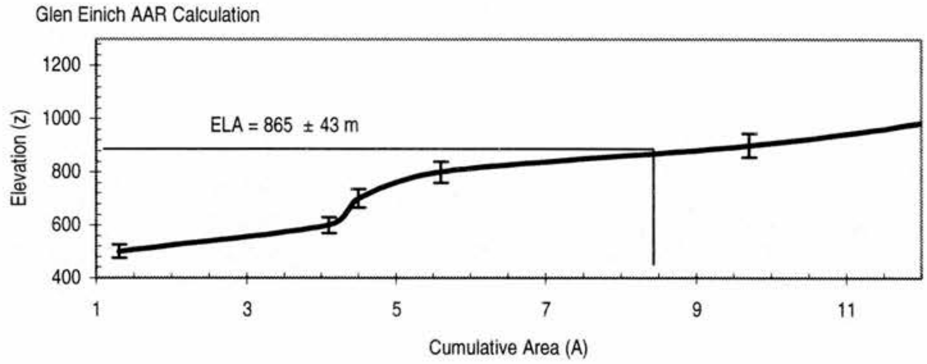


Figure 6.2: The Western Cairngorm Ice Cap with proposed Heinrich-1 correlative ELAs reconstructed, using an AAR value of 0.6 ± 0.05, and accumulation areas determined by ice divides.

a:



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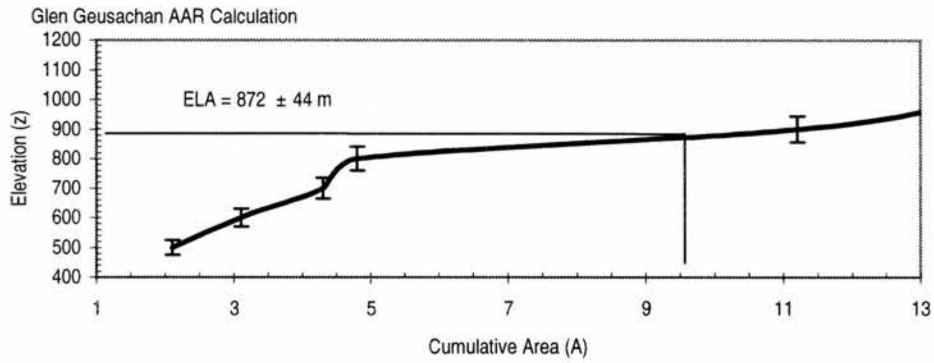

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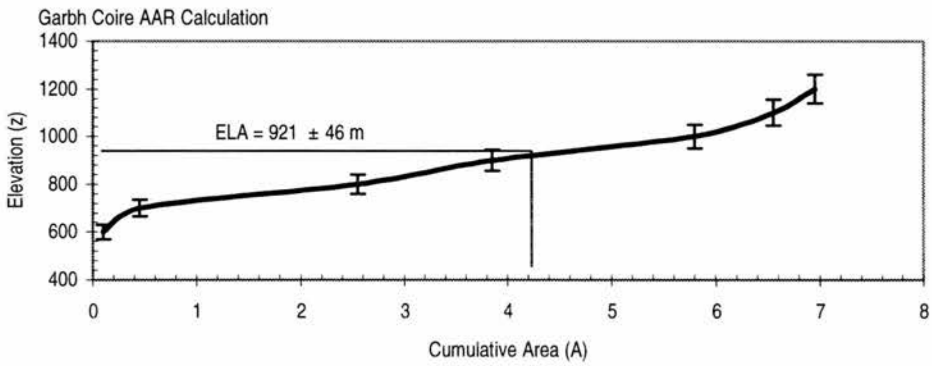

Figures 6.3a, b and c: Reconstructed ELA values for the Einich, Geusachan and Garbh Coire glaciers, using the AAR method.

ages derived from the glen. Once erosion and snow shielding corrections are included, then these ages fall outside the Stadial period. Zero erosion error weighted mean ages for the Glen Geusachan moraines argue for the start of the final phase of deglaciation of this part of the Cairngorms at 13,188 ± 529 BP, and once 10 mm / ka erosion corrections are completed then the age increases to 15,035 ± 605 BP.

The two moraines sampled are part of a suite of such features common in modern glaciated regions, indicative of steady retreat of an ice front. As such they represent a stage in the decline of a much larger ice mass, rather than the outermost extent of a glacial regrowth following total disappearance of ice at the end of the LGM. Similarly there is no indication in the landscape within these limits that such a regrowth ever occurred as argued by Sugden (1970). The features argue for the steady decline and retreat of the ice fronts, both north, south and east of Glen Geusachan in Glen Dee, followed by the final collapse of the ice mass.

Therefore earlier arguments concerning the occupation of Glen Geusachan by Loch Lomond Stadial ice (Sissons, 1979, Bennett & Glasser, 1991) cannot be upheld. The data presented here places the final large phase of glaciation of the glen earlier than the Loch Lomond Stadial. It is proposed that glaciation related to this cold phase was limited to the high corries, as proposed by Sugden (1968) and others (Purves *et al.* 1999). The Cairngorm Ice Cap did not disappear as a result of the rapid warming of 7° C over 50 years at the end of the Younger Dryas (Dansgaard *et al.* 1989). Rather it declined much earlier in the late glacial.

6.4: The 'hummocky moraine' debate

Following the reinterpretation of the Glen Geusachan glacial limits as belonging to the a period during the main deglaciation of the Cairngorms prior to the Loch Lomond Stadial, other previous interpretations relating to the particular geomorphology of the glen must be questioned. 'Hummocky moraine' has been used by many as an indicator of a Loch Lomond Stadial or Younger Dryas age of glacial deposits (Sissons, 1967, 1974, 1977, 1979, Benn, 1992, 1993, Bennett and Boulton, 1993a 1993b, Bennett 1991). While in many cases the moraines described may well have been deposited during the Stadial, the dating of the Glen Geusachan moraines to a pre Loch Lomond Stadial period means that no direct conclusions regarding age of features may be drawn, simply by analysis of their morphology.

The key to the 'hummocky moraine' debate lies within the association of these features with a particular climatic state. Rapid deglaciation at the end of the Loch Lomond Stadial has led to the formation of several groups of theories. The first proposed was that of stagnation, and downwasting in situ (Harker, 1901, Sissons 1967, 1974, 1977, 1979, Sugden, 1974, Clapperton and Sugden, 1977, Benn 1992, 1993). The second that active, rapid retreat caused 'hummocky moraine' formation at glacier margins (Charlesworth, 1956).

The final group of theories diverge from the climatically driven model of 'hummocky moraine' formation, and focus on glacial processes. Englacial thrusting at glacier margins has been proposed by Bennett et al, (1998), likewise the survival of discrete blocks of glacial ice buried beneath debris has received recent re-examination (Eyles, 1979, 1983, Everest and Bradwell, *in press*). Wilson and Evans (2000) argue that the term 'hummocky moraine' has been 'employed both descriptively and genetically to classify a range of landforms produced by a variety of glacier processes' (Wilson and Evans, 2000, p.149). The data from Glen Geusachan supports this view, in that the moraines here date from the lateglacial deglaciation of Scotland. There can be no strong link proposed between the 'hummocky moraine' in the Glen, and the climatic regime at the end of the Loch Lomond Stadial.

If this is the case, then areas that support 'hummocky moraine' outside Glen Geusachan that have been assigned Loch Lomond Stadial ages, may not have been glaciated during that period at all. Areas include the Isle of Skye (Benn, 1992, 1993) Coire a' Cheud-chnoic in Glen Torridon (Wilson and Evans, Bennett and Boulton, 1993, Bennett *et al*, 1998) Glen Callater (Sissons & Sutherland, 1976), and much of the northern extent of the proposed Loch Lomond Stadial ice sheet (Bennett and Boulton, 1993). This may provoke a reassessment of the extent of Younger Dryas glaciation in Britain, forcing a reduction in the accepted size of the glacial systems thought to have been present.

6.5: Correlation of Scottish Glacial Limits

The dated limit provided by the Glen More moraine damming the mouth of Glen Einich has wider implications in terms of the retreat of the main Scottish Ice Sheet (Figure 6.3). The low altitude of the margin implies that the Spey Valley glacier must have terminated within a few km of Glen More. This position is some 150 km within the proposed maximum offshore limit of the Scottish Ice Sheet at glacial maximum at the Bosies Bank Moraine. It thus illustrates a retreat of around 60% of the total retreat from offshore limits to the ice divide in the western Highlands by Heinrich-1 times.

It is possible that one or more of the readvances identified by earlier authors, such as the Wester Ross Readvance (Hinxman & Anderson, 1915, Charlesworth, 1955, Sissons, 1964, 1967, Paterson, 1974, Sissons & Dawson, 1981, Robinson & Ballantyne, 1979) correlates directly with the 16 – 15 ka stillstand identified here in the Cairngorms. The Wester Ross limit also lies some 150 km within west coast LGM limits, and also illustrates a retreat of 60% from maximum limits to the ice divide in the Western Highlands. This Wester Ross limit has not been dated, however it does lie outside proposed Loch Lomond Stadial moraine limits, and has been tentatively linked to H-1 by Benn (1997). The coincidence is interesting and could well imply that these two limits are contemporaneous. If ice flowing in the Spey began its retreat from glacial maximum positions at or around Bosie's Bank Moraine after 18 ka, then it took approximately 2 - 3 ka to retreat to the Cairngorms. Following one to three stillstands, the decline of the ice sheet was then far more rapid, with complete disappearance of ice from Scotland by around 14.5 ka. This represents an almost instantaneous retreat from the Glen More basin to the ice divide in the west.

Figure 6.3: The reconstructed western Cairngorm Ice Cap in relation to potential correlative limits, the Wester Ross Readvance limit (Robinson and Ballantyne, 1979, Sissons and Dawson, 1979, Benn, 1997), the Perth Readvance limit (Sissons, 1967, Sutherland, 1981), and shoreline limits at Otter Ferry, Glendaruel, Loch Long, Arisaig and Loch Morar (Sutherland, 1981).

Summary and Conclusions

There are several direct conclusions that can be drawn from this study.

1. For the first time a moraine sequence in the UK has been directly dated using cosmogenic surface exposure dating, constrained by ^{14}C.

2. A lateglacial stillstand has been identified that affected both Cairngorm ice and in the Scottish Ice Sheet between 16 and 15 ka BP. This stillstand lasted approximately 1 ka.

3. The stillstand correlates with Heinrich Event 1 in the North Atlantic. ELA calculations suggest a temperature decline of the order of 3°C compared to present day values. The stillstand may also correlate with regional moraine records within Scotland, such as those identified as the Wester Ross Readvance.

4. There is no evidence of widespread valley glaciation in the Cairngorms during the Loch Lomond Stadial. The 'hummocky moraine' often used to tie glacial deposits to this period relates to the 16 – 15 ka stillstand.

5. There was a period during the deglaciation of Scotland where high massifs, such as the Cairngorms, supported independent ice caps.

6. Cosmogenic Surface Exposure Dating of morainic boulders is an applicable technique to the establishment of the age and character of deglaciation in Scotland. The application of the cosmogenic surface exposure dating technique using the ^{10}Be radionuclide, is a robust and tested method. Nevertheless this thesis has demonstrated the importance of establishing field relationships of individual samples and their broader geomorphological context. This study also illustrates the need for cosmogenic surface exposure dating studies to be combined with other environmental dating techniques. This is important when the erosion or burial histories of cosmogenic samples are unknown and unknowable.

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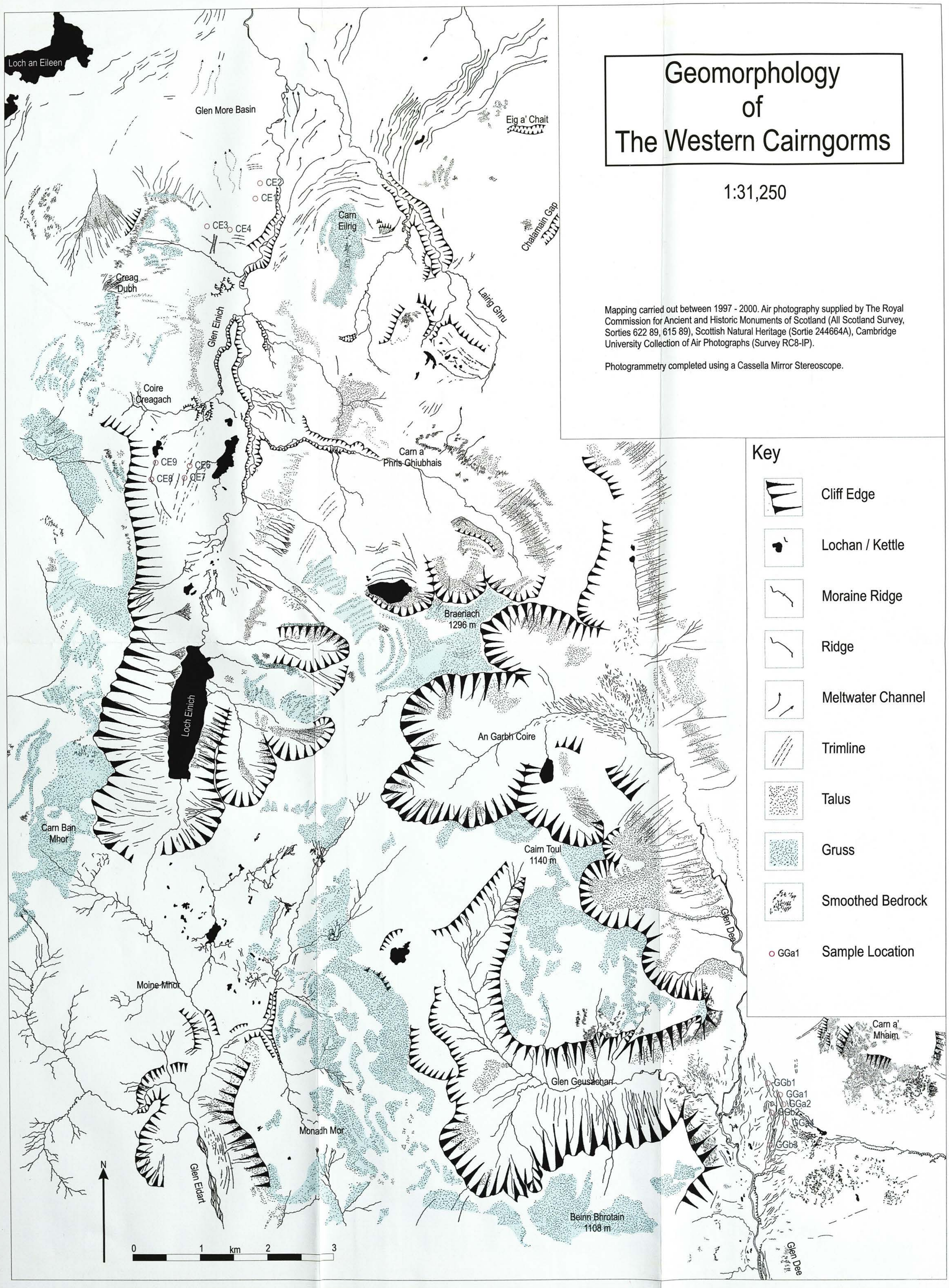
Geomorphology
of
The Western Cairngorms
1:31,250
Mapping carried out between 1997 - 2000. Air photography supplied by The Royal Commission for Ancient and Historic Monuments of Scotland (All Scotland Survey, Sorties 622 89, 615 89), Scottish Natural Heritage (Sortie 244664A), Cambridge University Collection of Air Photographs (Survey RC8-IP).
Photogrammetry completed using a Cassella Mirror Stereoscope.
Key
Cliff Edge
Lochan / Kettle
Moraine Ridge
Ridge
Meltwater Channel
Trimline
Talus
Gruss
Smoothed Bedrock
GGa1
Sample Location
Loch an Eileen
Glen More Basin
Eig a' Chait
Chalamain Gap
CE2
CE1
CE3
CE4
Carn Eilrig
Creag Dubh
Glen Einich
Lairig Ghru
Coire Creagach
CE9
CE6
CE8
CE7
Carn a' Phris Ghiubhais
Braeriach
1296 m
Loch Einich
An Garbh Coire
Carn Ban Mhor
Cairn Toul
1140 m
Glen Dee
Moine Mhor
Carn a' Mhaim
Glen Geusachan
GGb1
GGa1
GGa2
GGb2
GGa4
GGb3
Monadh Mor
Glen Eidart
Beinn Bhrotain
1108 m
Glen Dee
N
0
1
km
2
3